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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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QUATERNARY STRATIGRAPHY OF THE GUADIANA VALLEY, DURANGO, MEXICO

BY CLAUDE C. ALBRITTON, JR.

ABSTRACT

Interbedded lavas and sedimentary rocks in the Guadiana Valley are divisible into six units. The oldest consists of gravel which caps remnants of a pediment. An ancient canyon of the Río Mezquital, entrenched about 200 feet below this pediment, was filled with basalt, and the basalt in turn overlapped by gravelly alluvium of the Guadiana formation.

The Guadiana formation contains fossils of pre-Wisconsin Pleistocene age. After this alluvium accumulated, it was eroded and deeply weathered, first under dry and later under humid conditions, as indicated by superposed pedocal and pedalfer soils.

Following renewed eruptions of basalt, the Guadiana formation was covered in most places by humic silty alluvium of the Pueblito formation. Both the lava, as indicated by its freshness, and the Pueblito, as indicated by its fossils, are of Recent age. A humic zone at the top of the Pueblito contains hearths and artifacts identified with the Southwestern Desert Cultures, which were best developed between 6000 and 500 B.C.

Gullies cut into the Pueblito formation were later filled with sand and gravel. Prior to the cutting of the present arroyos, a thin blanket of silt, which contains Indian pottery and relics of Spanish and Mexican occupation, was spread over the lower parts of the Guadiana Valley. Some of the pottery corresponds with a Chalchihuites Culture occupancy dated *circa* A.D. 500–1350.

Tentatively the Guadiana formation, Pueblito formation, and the pottery-bearing silt are correlated respectively with the Lower Becerra, Totolzingo, and Nochebuena units of the Valley of Mexico, and with the Neville, Calamity, and Kokernot formations of western Texas. Units older than the Guadiana are included in the Quaternary because they are younger than an erosional surface of probable Pleistocene age developed upon the Sierra Madre Occidental.

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Introduction

Fire hearths and stone artifacts are occasionally found weathering from alluvium in the broad intermontane valleys of southern Durango. On the supposition that some of these objects might represent fairly ancient human cultures, Dr. J. Charles Kelley recently proposed that geologists and archeologists should study the Quaternary stratigraphy together.

The summer of 1956 offered an opportunity for this kind of an investigation. Dr. Kelley was then making his third expedition into Durango as director of the anthropological field session of Southern Illinois University. With the aid of a grant from the Penrose Bequest of The Geological Society of America, the author worked for a month with the archeologists. During this time a sequence of alluvial and lacustrine deposits, ancient soils, and lava flows was established. The stratigraphic horizons of buried cultural materials were then determined with reference to this sequence.

Prior to the geologic studies, Mr. James A. Pitkin scouted the country around the city of Durango and located the best exposures of the younger rock units on a route map prepared with odometer and compass. During a week of reconnaissance, Dr. Kelley led the party to archeological sites illustrating a succession of cultural horizons. Agnes McClain Howard of Durango City, after contributing the better part of her collection of vertebrate fossils, guided the group to localities where additional collections were made. Dr. James H. Quinn of the University of Arkansas has since identified the Howard collection, and Dr. E. P. Cheatum of Southern Methodist University has identified snail shells associated with some of the vertebrate fossils. In reviewing the manuscript for this report, Messrs. Carl Fries, Robert E. King, and J. Charles Kelley have offered many thoughtful and constructive suggestions.

Geography

The Guadiana Valley is a high plain bordering the eastern flank of the Sierra Madre Occidental in the southern part of Durango. The valley measures about 2100 square miles and has an average altitude of about 6200 feet. Although irregular in plan, it is elongated toward the northwest (Fig. 1). At the northwestern end is a bolson containing a large intermittent lake, the Laguna de Santiaguillo. Through a narrow pass, which marks an inconspicuous divide in the drainage, the bolson connects southward with the broader part of the valley, to which the name Guadiana is restricted by some authors. This is encircled by mountains, except at the north, where passes connect with the valley of the Río Nazas, and at the south along the valley of the Río Mezquital.

Three physiographic provinces meet in this area (Fig. 2, index map). The highland west of the Guadiana Valley is part of the Sierra Madre Occidental, whereas that to the south belongs to the Mesa Central. The valley itself and the ranges that rise north and east of it are parts of the Mesa del Norte (Garfias and Chapin, 1949, Fig. 1, p. 21–26, 42, 69–70).

In Durango, as in Chihuahua to the north, the Sierra Madre Occidental is

> ..."a great plateau, fringed by mountains on the east, trenched by deep cañons in its center and bordered by a wild and rugged complex of mountains carved out of the plateau on the west" (Weed, 1902, p. 444).

Near the 24th parallel the ascent from the Guadiana Valley to the surface of the plateau is over foothills a few hundred feet in height. Thence westward for about 45 miles, the plateau rises on an average of 50 feet to the mile to crests at elevations of 8800 feet or more, beyond which the slopes break sharply down toward the Pacific Ocean. At higher latitudes the relief along the border between the Sierra Madre and the Guadiana Valley is commonly 1000 feet or more, and the transverse profile of the plateau is correspondingly flatter.

In the vicinity of the Guadiana Valley, the Sierra Madre and the Mesa Central are not sharply divided. Both provinces are plateaus, but the Sierra Madre is in general higher and more intricately dissected, whereas the contours of the Mesa Central are modeled to a greater extent by outpourings of lava (Garfias and Chapin, 1949, p. 69–70).

The Mesa del Norte is a high tableland characterized by alternating sierras and broad valleys. In southern Durango the lowland portions stand a mile or more above sea level, although their elevations decrease toward the north. Most of the sierras trend between north and northwest. Several of the larger ranges rise 2000 to 3200 feet above the surrounding plains.

The climate of the Guadiana Valley is semiarid. Between 16 and 24 inches of rain falls during the year, mostly in July, August, and

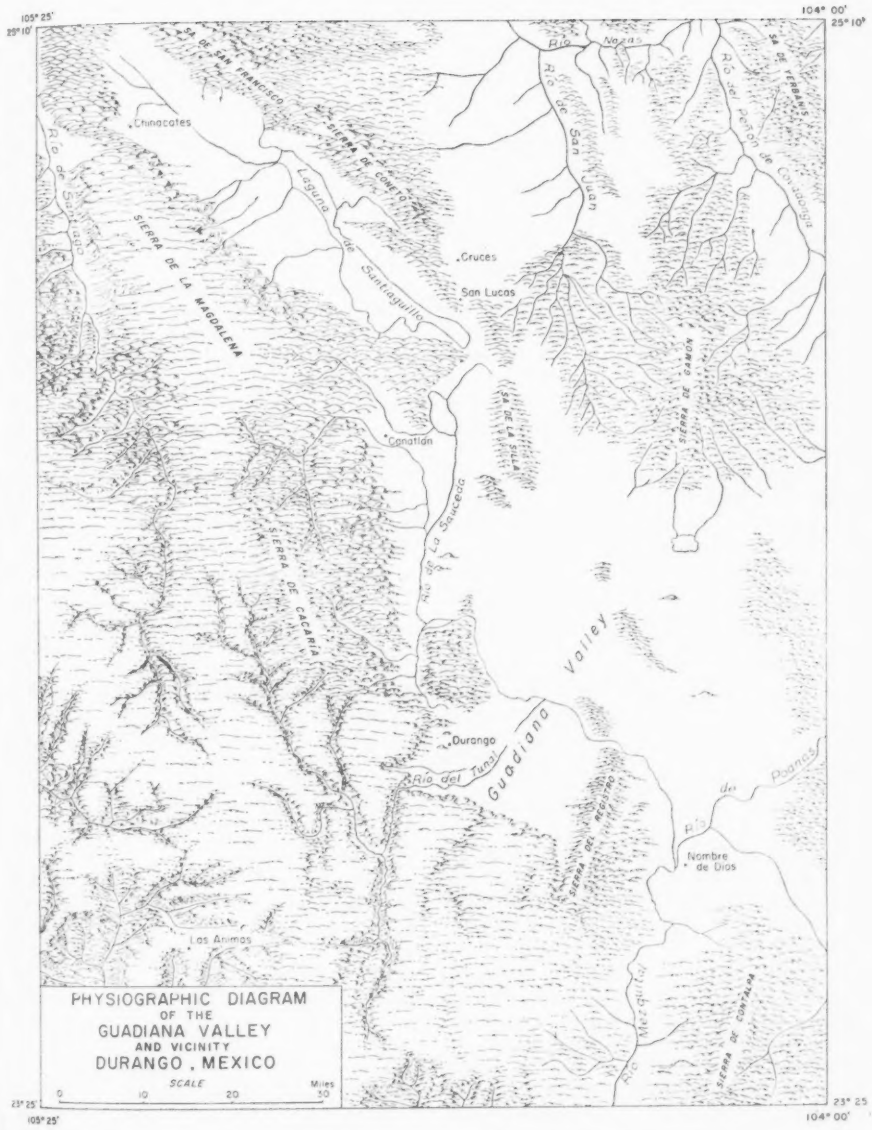


FIGURE 1.—PHYSIOGRAPHIC DIAGRAM OF THE GUADIANA VALLEY AND VICINITY, DURANGO, MEXICO

September. Summers are warm, winters mild: the mean monthly temperature for January is 50° F., whereas from April through October the monthly means range between 60° and 70° F. (Vivó and Gómez, 1946). Rainfall increases westward with the increase in altitude; the crest of the Sierra Madre receives 30–40 inches in normal years (Vivó and Gómez, 1946; Tamayo, 1949, p. 7 of Atlas; De la O Carreño, 1950).

The area in and around the Guadiana Valley drains partly to bolsons on the Mesa del Norte

and partly to the Pacific Ocean. The Laguna de Santiaguillo, which was practically dry in 1956, occupies the lower part of an independent drainage basin. North-flowing streams on either side enter the Río Nazas, which in turn ends at a bolson near the city of Torreón (*see* Tamayo, 1949, Atlas, p. 10). The part of the Guadiana Valley directly south of the Sierra de Gamón is for the most part a rough field of lava, upon which no integrated pattern of drainage is developed. Río de la Sauceda, a perennial stream heading in the mountainous border of the Sierra Madre, skirts the southern edge of the lava, joins the Río del Tunal, and drains to the Pacific through the Mezquital. Streams draining the higher slopes of the Sierra Madre, west of the headwaters of the Tunal, reach the ocean by less circuitous courses through gorges cut into the steep western front of these mountains.

An open forest of long-leafed pine once covered the higher parts of the Sierra Madre. The cutting of trees for timber and charcoal, however, has substantially reduced the forest cover and may explain the prevailingly treeless character of uplands adjacent to the Guadiana Valley.

Uncultivated alluvial stretches of the Guadiana Valley are commonly grassy or covered with mesquite. Poplar and willow grow along the water-courses, and cypress thrives in marshy reaches of the Mezquital, where it flows around the northern end of the Sierra del Registro. Rocky slopes around the edges of the valley support thick stands of the giant prickly pear.

The city of Durango, with a population of more than 75,000, is the capital of the state and principal city of the region. From the Cerro de Mercado, the famous "mountain of iron" which rises from the Guadiana Valley near the northern outskirts of the city, comes the greater part of Mexico's high-grade iron ore (Foshag, 1929; Flores, 1950). Railways connect Durango with mining and lumbering camps in mountains to the north and west, as well as with the principal cities of central Mexico. Highways radiate in all directions except south, and a network of farm roads provides easy access to most parts of the countryside.

Geologic Setting

Volcanic Rocks of the Mountains

The Sierra Madre Occidental is made largely of Cenozoic volcanic rocks. In parts of Durango and Chihuahua the older part of the volcanic sequence is dominantly andesitic, and the younger part is rhyolitic (Fowler, 1900; Weed, 1902; Terrones Benítez, 1922; Santillán, 1929). The total thickness of the volcanic rocks is estimated to be on the order of 5000 to 6000 feet (Singewald, 1936, p. 1163–1170; Imlay, 1939, p. 1725–1740).

Sierras of the Mesa del Norte that border the western Sierra Madre are made of these same volcanic materials. Here also the lower parts are commonly andesitic and the upper parts rhyolitic, as in the Mapimí area of northeastern Durango (Singewald, 1936, p. 1163–1170) and in the Sierra de San Francisco farther south (Foshag and Fries, 1942, p. 170–176; Smith, Segerstrom, and Guiza, 1950, p. 171–173).

In neither province, however, is the age of the volcanic sequence well established. King (1939, p. 1715–1717) classified the volcanic rocks of the northern Sierra Madre Occidental as early Tertiary, and Edwards (1955, p. 182–183) has paleontological evidence to show that the beginnings of volcanism, at least on parts of the Mesa Central and Mesa del Norte, antedate the late Eocene. At the other extreme, Burrows (1953) would have all the volcanic accumulations of the western Sierra Madre, along and near the 24th parallel, no older than Pliocene; Fowler (1900) suggested that the younger part of the rhyolitic series may range into the Quaternary. Perhaps, as Garfias and Chapin (1949, p. 51–53) have suggested, there has been intermittent volcanism in western Mexico from the beginning of the Tertiary until the present.

The highlands around the Guadiana Valley are largely made of pyroclastic rocks belonging to the younger rhyolitic sequence, which is probably at least 3200 feet thick in the Sierra de San Francisco (Smith, Segerstrom, and Guiza, 1950, p. 171–173). The tuffaceous rocks are commonly stratified in layers ranging from a few inches to several feet thick. In the foothills around the city of Durango, fine-textured, light-gray and pinkish tuffs are quarried for building stone. Thin sections of samples taken from quarries near Pueblito showed that on an average about one-tenth of the rock is made of fresh crystals and fragments of orthoclase and sodic plagioclase set in a weakly birefringent ashy matrix. Tuffs of coarser texture, made largely of crystals, appear at various horizons in the sequence, commonly as massive units several score feet in thickness. Specimens from cliffy slopes near Locality 13 (Fig. 2) consist

predominantly of sanidine fragments up to 3 mm across, packed in a fine-textured matrix containing bits of partly devitrified pumice (*see also* Salazar Salinas *et al.*, 1923, p. 65).

Numerous authors have described lava flows interbedded with the pyroclastic rocks. For example, Foshag and Fries (1942, p. 165–167) mention evenly layered flows of rhyolite at

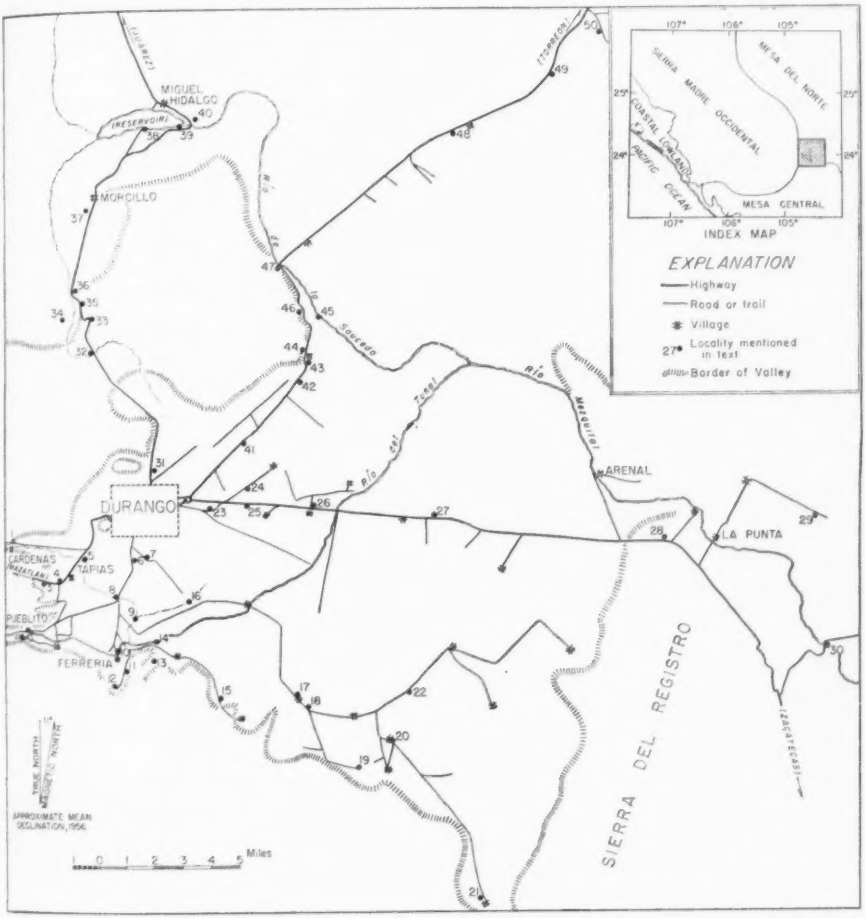


FIGURE 2.—ROAD GUIDE TO EXPOSURES OF QUATERNARY ROCKS AROUND DURANGO CITY
Map by James A. Pitkin, 1956. Distances measured by odometer; bearings by compass.

Parts of the tuffaceous sequences are evenly bedded and contain lentils of well-rounded gravel (Locality 37, Fig. 2). These accumulations must have been deposited by water on broad flood plains or in lakes. Singewald (1936, p. 1163–1170) has described water-laid sediments of pyroclastic origin in the volcanic complex of northeastern Durango. In the northwestern part of the state, Pérez Siliceo and Gallagher (1947) also found stratified tuff that is obviously water-laid.

Cerro de los Remedios, a hill rising along the western outskirts of Durango City. Rhyolite and latite are known from the Cerro de Mercado directly north of the city (Flores, 1950, p. 112–115).

Possibly some or even most of the bodies described as rhyolite flows are either pyroclastic or intrusive. Smith, Segerstrom, and Guiza (1950, p. 193–196) suggest that some of the supposed rhyolitic flows of the Sierra de Cacaría may be welded tuffs. Burrows (1953)

believes that many of the so-called dacite, andesite, and rhyolite flows in the western Sierra Madre are actually laccoliths and sills intruding stratified pyroclastic strata.

Structure

In the vicinity of the Guadiana Valley the differences in physiography between the Sierra Madre and the Mesa del Norte reflect fundamental differences in structure. Directly west of Durango City the tuffaceous strata of the Sierra Madre are in general inclined toward the east at angles averaging less than 5°. Along the 24th parallel this regional dip persists for several tens of miles to the crest of the mountains, so that this part of the range has the form of a great cuesta. By contrast, the prevailing dip of the volcanic rocks in mountains south of the city of Durango is northward at an average of about 5° (Pl. 1, fig. 2). The change in strike seems to coincide with the long southern tributary of the Río del Tunal, which is probably a subsequent stream following a zone of north-trending fractures (Fig. 1). This valley was not examined, but minor faults trending between northeast and northwest are common farther to the north in foothills west of Durango City. Between Durango and Chinacates, the bold eastern front along the Sierra de la Magdalena and the mountains to the south appears to be a fault escarpment.

Most of the mountains east of the Sierra Madre foothill belt have the elongate plan and asymmetric profiles suggestive of fault blocks. The volcanic rocks of the Sierra del Registro have a general northerly strike and dip 15°–30° eastward. The strike is almost at right angles to that of the rocks in mountains south of Durango City, and the field relationships suggest that a second fault zone borders the western front of this range. Little is known of the structure of mountains northeast of the Laguna de Santiaguillo, but the alignment of their bases parallel with the escarpment of the Sierra de la Magdalena again suggests faulting (Fig. 1).

Burrows (1953) pictures the Sierra Madre Occidental as a gigantic horst, uplifted along longitudinal fractures above the Mesa del Norte on the east, and to a far greater degree above the Gulf of California on the west. Granting that the eastern border follows a fault zone, it is not necessary to conclude that its eastern side, topographically the lower, was downthrown. Low hills north of Durango City contain andesite (Locality 46, Fig. 2), a rock more commonly associated with the older than the younger volcanic sequence in the Sierra Madre. These hills also contain more conglomerate than is common in the younger rhyolitic sequence of these parts. The lithology of a similar group of hills north of Canatlán was not studied, but where these hills connect with the Sierra de la Magdalena, the imposing eastern escarpment of this range disappears, suggesting that the scarp is not an initial feature of the landscape but rather a fault-line scarp produced by differential erosion. Foshag and Fries (1942, p. 170–176) found pre-Tertiary rocks beneath the older Tertiary andesite of the Sierra de San Francisco, from which it would appear that this range has been uplifted rather than depressed with relation to rhyolitic sequences a few miles westward across the Santiaguillo bolson.

An Ancient Surface of Erosion

The plateau of the Sierra Madre west of Durango City is an ancient erosional surface trenched by canyons of more recent origin. Residuals of this surface are smoothly contoured, with scattered irregular clusters of hills rising a few hundred feet above broad valleys and rolling plains (Pl. 1, fig. 1). The plateau slopes eastward at angles less than 1°. Although this slope locally coincides with the dip of underlying volcanic rocks, it generally truncates these strata at low angles. What is evidently a northern continuation of this same surface is extensively developed in the states of Chihuahua and Sonora (King, 1939, p. 1627, 1699, 1715–1717).

Along the highway between Durango and Mazatlán the upland surface bears no capping of fluviatile gravel but is mantled by residual soil. Rounded stream gravel scattered alongside some stretches of the highway was hauled from near-by stream terraces.

Many of the canyons in the plateau are at nearly right angles to the eastward slope of the old erosional surface (Fig. 1). Most canyons in the headwaters of the Santiago, Sauceda, and Tunal trend north to north-northwest, as though they were following some regional pattern of fracture. In advance of the detailed mapping required to settle the issue, it may be supposed that these are subsequent streams which may have little in common with the drainage responsible for the ancient surface of erosion.

Profound changes in the drainage around the Guadiana Valley, incident to the cutting of the

present canyons, are suggested not only by the pattern of the streams but also by the character of the fish fauna. Of 11 species collected by Meek (1904, p. xxvii; xxxiii–xxxvii) from the Río del Tunal near the city of Durango, 8 belonged to the present fauna of the Rio Grande. The Rio Grande fauna is also well represented in the Río Nazas and other streams of the Mesa del Norte. Meek could find no alternative to the proposition that the headwaters of the Mezquital once connected northward with the Rio Grande, even though the two rivers are now separated by 350 miles or more of semidesert.

Recent investigations of the alluvial deposits along the Rio Grande (Smith and Albritton, in press) indicate that the part of the river flowing through the El Paso Valley, in the extreme western tip of Texas, came into existence during the Nebraskan stage of the Pleistocene. If the upper part of the river is no older than this, then the migration of Rio Grande fishes into the Tunal drainage likewise must have occurred during the Quaternary, and the capture of the Tunal drainage by the Mezquital must have followed later in this period.

Thus it may be assumed that following eruption of the rhyolitic volcanic rocks, the eastern Sierra Madre was reduced to a surface of low relief, merging, probably, with intermontane lowlands of the Mesa del Norte, and draining north and east into the Rio Grande during the early part of the Ice Age. As a result of regional uplift during the Pleistocene (Burrows, 1953), streams draining the western slopes of the mother range were rejuvenated. In time these cut headward to capture headwater branches of the Rio Grande, many of which had adjusted themselves as subsequent streams along meridional lines of fracture.

Rocks of Quaternary Age

General

The rocks here referred to the Quaternary system consist of pediment gravel, lava, alluvium, coluvium, lacustrine marl, aeolian deposits, and buried soils. The six principal units are shown in the columnar section (Fig. 3). Together these form a thin but complex blanket which covers the lower parts of the Guadiana Valley and extends short distances up the slopes of mountains rising within and around it.

Pediment Gravel

The gentler eastern slope of the Sierra del Registro is a dissected pediment slanting toward the valley of the Mezquital. The highway connecting the city of Durango with Nombre de Dios skirts the lower slopes of the pediment, and the cuts along the highway expose gravel deposits beneath its surface (Fig. 4).

The gravel is made largely of cobbles and boulders derived from rhyolitic rocks of the Sierra del Registro. Around residual hillocks and ridges of bedrock rising above the pediment, the gravel is no more than a few feet thick, but beneath the broader expanses of the pediment it may be several tens of feet thick. Most of the stones are subangular and set in a matrix of pebbly tuffaceous sand. Thick crusts of caliche are common in the upper parts of the gravel, but in many places the caliche has been leached from the upper foot or so. The zone of leaching is reddish-brown, and although the stones within it remain intact, the matrix is partly altered to clay.

Spurs capped with the pediment gravel stand as high as 200 feet above the bed of the Mezquital where the river flows in a canyon a short distance southeast of Locality 30. Evidently the pediment was originally graded to this same valley prior to the cutting of the canyon.

A small inlier of similarly coarse gravel is exposed in the outlet of the reservoir at Miguel Hidalgo (Locality 40, Fig. 2). Here boulders of crystal tuff fill a channel 13 feet deep and 35 feet wide that was cut into the tuffaceous bedrock and later buried by lava (Pl. 2, fig. 1). Exposures are too limited to show whether the gravel is related to a pediment developed upon hillsides near by, but the situation suggests this relationship.

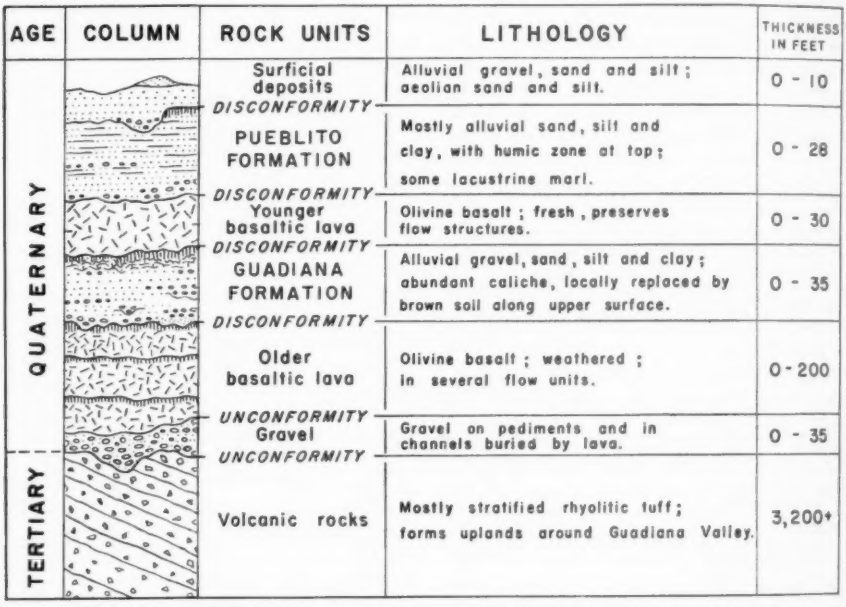
Although it is not established that the pediment gravel of the Sierra del Registro and the channel fill of Miguel Hidalgo are of the same age, both deposits are younger than the upland surface of the Sierra Madre: both are related to drainage lines that lie well below the upland level projected eastward over the Guadiana Valley. If the upland originally drained to the Rio Grande, and if the Rio Grande in its upper reaches is a Quaternary stream, the gravel must also be Quaternary.

Older Basaltic Lava

About half the Guadiana Valley is covered or underlain by basalt. The principal lava field measures between 900 and 1000 square miles.

It occupies an area bordered on the south by the Mezquital and Sauceda rivers and on the north by a line connecting Canatlán with the southern end of the Sierra de Gamón.

rocks that form the adjacent mountains (Fig. 4). This valley was about 1,25 miles across and 150–200 feet deep. Upstream from the falls, the lava fills the valley to its borders on either



AGE	ROCK UNITS	LITHOLOGY	THICKNESS IN FEET
QUATERNARY	Surficial deposits	Alluvial gravel, sand and silt; aeolian sand and silt.	0 - 10
	DISCONFORMITY		
	PUEBLITO FORMATION	Mostly alluvial sand, silt and clay, with humic zone at top; some lacustrine marl.	0 - 28
	DISCONFORMITY		
	Younger basaltic lava	Olivine basalt; fresh, preserves flow structures.	0 - 30
	DISCONFORMITY		
	GUADIANA FORMATION	Alluvial gravel, sand, silt and clay; abundant caliche, locally replaced by brown soil along upper surface.	0 - 35
	DISCONFORMITY		
	Older basaltic lava	Olivine basalt; weathered; in several flow units.	0 - 200
	UNCONFORMITY		
	Gravel	Gravel on pediments and in channels buried by lava.	0 - 35
	UNCONFORMITY		
TERTIARY	Volcanic rocks	Mostly stratified rhyolitic tuff; forms uplands around Guadiana Valley.	3,200+

FIGURE 3.—COLUMNAR SECTION OF STRATIFIED ROCKS IN THE GUADIANA VALLEY

Colluvial deposits not shown. Disconformities indicated above and below younger basaltic lava are based on physiographic evidence only. Short vertical lines, drawn along some of the contacts, indicate zones of weathering.

The lava field consists of many individual flows. Residents of this area draw a distinction between the rough and almost impassable country covered by the fresher flows (breña) and the more level terrain developed upon the older ones (malpais). This distinction has stratigraphic as well as physiographic significance, because the younger and older eruptions of lava in the Guadiana Valley were separated by an interval of alluviation that can be dated as Pleistocene on paleontologic evidence.

The older lava was studied at two localities near the southern edge of the lava field. Locality 30 (Fig. 2) marks the place where the Mezquital falls 115 feet over a cliff of basalt and enters a canyon cut into this same rock (Pl. 1, fig. 3). Locality 40 is on the edge of a ravine forming the outlet for the reservoir at Miguel Hidalgo.

Around the falls of the Mezquital, locally called El Saltito de la Constancia, the basalt fills an older valley cut into the same rhyolitic side. Less than a mile below the falls, most of the lava has been removed by erosion, and the valley has been re-excavated to near its original level.

The rock at El Saltito is olivine basalt in horizontal flow units averaging about 30 feet thick. Zones of weathering separate one unit from the next. A dark-gray amygdaloidal rock, the basalt is fine-textured and porphyritic. A thin section of a typical specimen shows phenocrysts of olivine, augite, and labradorite embedded at random in dark glass. Clusters of phenocrysts contain abundant intergranular olivine. Most of the feldspar is partly altered to clay minerals, and vesicles are filled with intergrowths of carbonate minerals and opal. Within the present zone of weathering, the lava is iron-stained, and most of the amygdules have disappeared.

At Miguel Hidalgo, the thin marginal portion of the lava rests upon shoulders of rhyolitic tuff that rise to connect with hillsides directly

southward. About 40 feet of vesicular olivine basalt is here exposed in the banks of the Sauceda, but on the spillway of the reservoir 100 yards to the south, the section is less than 10 feet thick (Pl. 2, fig. 1). This thinner portion consists of two superposed units, each about 3 feet in maximum thickness, with 2 feet of brownish-gray pebbly tuff at the base. Before it was buried, much of the older flow had weathered to spheroidal boulders.

A thin section of the younger flow at Miguel Hidalgo shows phenocrysts of olivine and labradorite set in dark glass. The glassy matrix contains randomly oriented microlites of plagioclase. Microlites and matrix together make up about half the solid portions of the rock. The glass is devitrified in spots, and many of the olivine crystals are partly altered to iddingsite.

The surface of the malpais consists of flat areas alternating with ridges and hillocks of boulders. The flats represent initial depressions in the lava field that have since become floored with sediment; the larger flat areas cover an acre or so of ground and are cultivated. The rocky ridges consist for the most part of residual boulders which mark the higher portions of the original lava field.

Processes responsible for leveling the surface of the older flows probably differed in kind from place to place. Parts of the lava field around the falls of the Mezquital preserve cappings of gravel which show that the basalt was buried by alluvium and later exposed by erosion (Fig. 4). In this situation the flat areas are essentially outliers of younger alluvium. Elsewhere, however, there is nothing to suggest that the lava has ever been buried, and in this instance the flats have probably been leveled after the manner of ephemeral lake beds, by accumulation of the finer products of weathering washed down from encircling ridges of higher ground, together with whatever windblown dust had settled between showers.

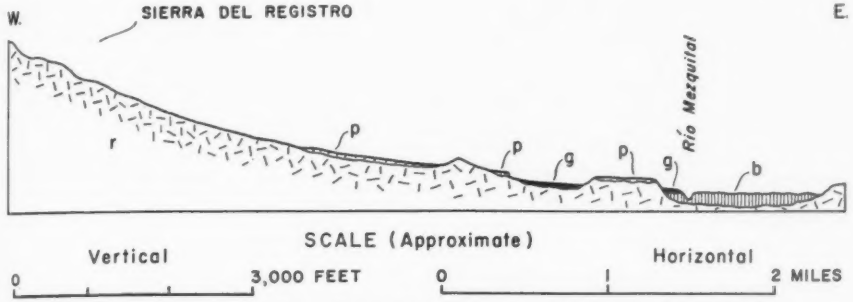


FIGURE 4.—DIAGRAMMATIC CROSS SECTION OF MEZQUITAL VALLEY AND EASTERN SLOPE OF SIERRA DEL REGISTRO

Through Locality 30 (Fig. 2). Showing relationships between rhyolitic bedrock (r), pediment gravel (p), older basalt flows (b), and Guadiana formation (g). Prepared from field sketches and photographs.

Guadiana Formation

General features.—The Guadiana formation is the name here given to a deposit of Pleistocene alluvium laid down within the Guadiana Valley and neighboring intermontane lowlands. It consists of gravel, sand, silt, and clay derived from rocks of the surrounding hills and mountains. Around the outer edges of the lowlands, the formation forms flat-topped spurs inclined gently away from the mountains. Along the axial parts of the lowlands, much of it has been destroyed by erosion; the remnants are buried beneath younger alluvium. Artificial exposures along roads or in gravel pits show, however, that this deposit must once have covered most or all of the Guadiana Valley as a blanket at least 30 feet thick in many places. Where the base of the alluvium is exposed, the lower parts in general rest unconformably on the Tertiary bedrock, but locally the formation disconformably overlies the older basalt of the malpais. Although fossils are not common, bones and teeth of vertebrate animals from one locality establish the age of the formation as Pleistocene, and suggest further that it is not younger than the Sangamon interglacial stage.

Aside from its stratigraphic position, the Guadiana formation may ordinarily be distinguished from younger alluvial deposits by its coarse gravelly or sandy texture, its light-

brownish color, and by the fact that it commonly contains conspicuous amounts of secondary carbonate deposits in the form of caliche.

Type locality.—The type locality is in a western re-entrant of the Guadiana Valley at the village of Cárdenas (Locality 1, Fig. 2). Here the formation makes a terrace at a level 16 feet above the flood plain and 31 feet above the bed of the arroyo that drains the re-entrant. The best exposures are in the northern bank of this arroyo, beginning at a point directly opposite the village and continuing several hundred yards downstream.

In its type area, the formation consists almost entirely of coarse gravel (Pl. 2, fig. 2). Cobbles and boulders of lava and tuff, derived from the bedrock of the surrounding hills, are the principal constituents. On the whole the gravel is poorly sorted: boulders up to 2 feet in diameter are scattered through zones in which the stones average 6 inches in diameter. Spaces between the larger fragments are filled with pebbles and granules, so that the aggregate is sufficiently compact to stand in steep banks. Regardless of size, the particles of tuff are well rounded, but most of the lava fragments are angular. The gravel is massive, except where interstratified with lentils of pebbly sandstone or disposed as cross-bedded layers filling depressions produced by scour. The maximum thickness of the gravel along the arroyo is 31 feet.

Other exposures in the Guadiana Valley.—The coarse gravelly facies of the formation is also seen in a roadside quarry at Locality 4, in an arroyo bank at Locality 21, and along the road over the malpais directly north of Locality 30. Bouldery slopes along piedmont benches at Locality 18 are also probably weathering from deposits of Guadiana gravel, although no exposures were found in this vicinity.

A mixed facies, consisting principally of interbedded cobbles, pebbles, and sand is exposed in a ditch along the Mazatlán Highway at Locality 3, as hummocky inliers along arroyo banks at Locality 5 and upsteam for about a mile, around the walls of large gravel pits at Locality 24, in road cuts at Locality 36, and in numerous gullies in the piedmont flanking the western slopes of the Sierra de la Silla.

A third facies, characterized by a predominance of silt and clay, is exposed alongside the road at Locality 7, in excavations around the abandoned bull ring (Locality 6), in pits at the city dump of Durango (Locality 31), in a drainage ditch at Locality 23, and in cuts along the Zacatecas Highway (Locality 26). South of the reservoir at Miguel Hidalgo, 5 feet of silty clay rests disconformably on basalt. The alluvial plain bordering the eastern front of the Sierra de la Magdalena around Chinacates and to the southeast is largely underlain by brownish silt and silty clay here referred to the Guadiana formation.

Not all exposures of the formation may be readily classified according to facies. Where the Guadiana Valley extends as embayments into the bordering mountains, the alluvium contains a mixture of finer material deposited by axial streams and of coarser material deposited by side streams. For example, in the

PLATE 1.—GUADIANA VALLEY AND SURROUNDING UPLANDS

FIGURE 1.—Sierra Madre Occidental, viewed northward from highway to Mazatlán, 12 miles southwest of Durango. Shows ancient erosional surface of low relief, gashed by headwater canyon of Río del Tunal. Residual hills rise several hundred feet above the general level of the plateau in right background. Elevation at rim of canyon is about 6500 feet.

FIGURE 2.—Southern border of Guadiana Valley, viewed toward southeast from near Ferrería. Sierra del Registro on far skyline. Hills to right are mostly of crystal tuff in even beds gently inclined northward toward valley. Floor of valley largely covered with friable silty alluvium of Recent age.

FIGURE 3.—El Saltito de la Constancia, the falls of the Río Mezquital (Locality 30, Fig. 2), viewed from the southeast. Falls developed upon horizontal layers of basalt filling old stream valley. Cuestas in background are foothills of Sierra del Registro.

PLATE 2.—OLDER BASALTIC LAVA, ALLUVIUM, AND BURIED SOILS

FIGURE 1.—Exposures in outlet of Río de la Sauceda below spillway of reservoir at Miguel Hidalgo, viewed southward from Locality 40 (Fig. 2). Boulder gravel (gr) fills channel cut in crystal tuff (t) and covered by basalt (b).

FIGURE 2.—Gravel of the Guadiana formation at the type locality (Locality 1, Fig. 2).

FIGURE 3.—Paleosols developed upon Guadiana formation; exposed in quarry at Locality 24 (Fig. 2). Much of the thick bed of caliche (c) developed in upper part of sand and gravel of Guadiana formation is leached and replaced by brownish ferruginous soil (f). Both paleosols buried by surficial layer of pebbly silt (s).

FIGURE 1

FIGURE 2

FIGURE 3

GUADIANA VALLEY AND SURROUNDING UPLANDS

FIGURE 1

FIGURE 2

FIGURE 3

OLDER BASALTIC LAVA, ALLUVIUM, AND BURIED SOILS

FIGURE 1

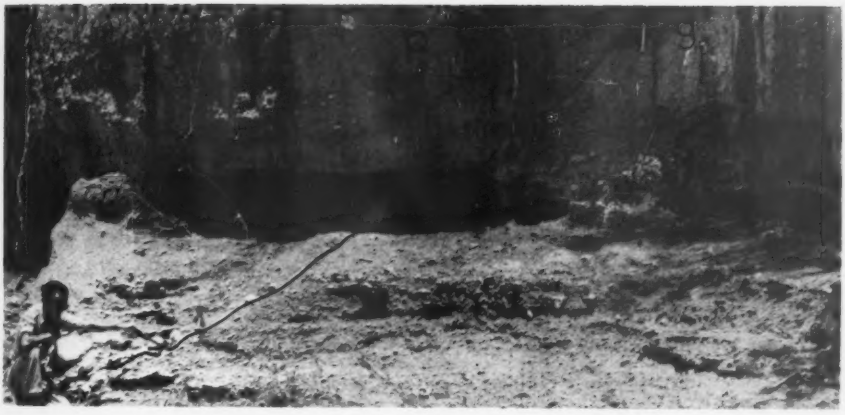


FIGURE 2

FIGURE 3

ALLUVIAL DEPOSITS

re-entrants around Ferrería (Localities 11, 12, 13) the alluvium is a complex mixture of interbedded gravel, sand, and silt, with no single grade class predominating.

Exposures in headwaters of Río de San Juan.—Where the alluvial plain along the southern end of the Laguna de Santiaguillo continues through a narrow constriction between hills east and west of San Lucas (Fig. 1), exposures along arroyos show that the alluvial formations of the Guadiana Valley continue into the interior drainage of the Río Nazas. In the banks of a small stream flowing through San Lucas, the Guadiana formation is represented by 3 feet of angular cobble and boulder gravel derived from and resting unconformably upon tilted beds of tuff. Farther north, where the formation is exposed in the banks of an arroyo at the small settlement of Cruces, it consists of interbedded sand and gravel 18 feet thick (Pl. 3, fig. 2).

Fossils.—From the gravel at Cruces, Agnes McClain Howard collected fragmentary bones of elephant and horse and assembled them in her private museum at Durango City. Guided by Mrs. Howard, the party collected additional material from this same locality. The choice items from the Howard museum and those more recently collected have been identified by Dr. James H. Quinn of the University of Arkansas.

Dr. Quinn recognized three species of vertebrate animals. A small equid is represented by half a lower jaw, an upper molar, the distal end of a femur, and a first phalanx. According to Dr. Quinn (personal communication, October 4, 1957) these fragments probably, although not necessarily, represent a single species, in which case the animal most closely resembles the European species *Asinus hydruntinus*. A larger horse (*Equus*), represented by the distal end of the tibia, calcaneum, astragalus, the proximal end of the metatarsal bone, and a first phalanx, cannot be identified according to species. *Testudo*, a large land turtle, is represented by a peripheral plate.

Age.—Dr. Quinn can find no record of *Testudo*, at least for this region, in deposits younger than the Sangamon stage of the Pleistocene. This same genus of turtle has been found in the lower member of the Becerra formation (Becerra Inferior) near Tequixquiac in the vicinity of Mexico City. Partly on this evidence Hibbard (1955, p. 49, 82) also dated the Lower Becerra as pre-Wisconsin.

Caliche

At some time following deposition of the Guadiana formation, large amounts of calcium carbonate were added to this body of alluvium. Within the silty and fine sandy facies the carbonate mineral commonly forms discrete pellets or irregular nodules which give the rock a spotted appearance. In accumulations of coarse gravel many of the cobbles and boulders are covered with rinds of the caliche. Massive beds of white caliche are characteristic of the upper few feet of mixed pebble gravel and sand.

A massive cap of caliche as much as 5 feet thick is well exposed around the edges of gravel pits at Locality 24 (Pl. 2, fig. 3). The caliche is also exposed at many places in ditches along highways leading east and northeast from Durango City, as for example at Localities 25 and 26 on the road to Zacatecas and at Localities 41 and 42 on the way to Torreón. Between Localities 47 and 49 on the Torreón highway the caliche crust is in most places only a few inches below the surface. At San Lucas (Fig. 1) 1 foot of unusually hard caliche at the top of the Guadiana formation stands in relief as a ledge along the arroyos.

Crusts of caliche surrounding cobbles and boulders are exposed in a gully along the Mazatlán Highway at Locality 3 and in a roadside quarry at Locality 4.

Disseminated nodules and pellets of caliche in silty phases of the formation may be seen around Durango City in excavations along a drainage ditch at Locality 23, around the abandoned bull ring at Locality 6, and at the garbage disposal dump (Locality 31).

Caliche of the kind present in the Guadiana formation is in general regarded as an accumu-

PLATE 3.—ALLUVIAL DEPOSITS

FIGURE 1.—Exposure in roadside gully 10 miles south of San Lucas on Durango-Juárez highway. Guadiana formation (g) at base of gully is whitish because of abundant caliche. Surficial silt, sand, and gravel (s) fill channel cut into gray clayey silt of Pueblito formation (p).

FIGURE 2.—Pueblito formation (p) filling channel cut into Guadiana formation (g); exposed in arroyo at Cruces.

FIGURE 3.—Interbedded gravel, sand, and silt of Pueblito formation exposed in bank of Río del Tunal at Pueblito, the type locality.

lation in the lower parts of soils and weathering profiles developed during protracted stages of dry climate. The concentration of the caliche toward the top of the Guadiana formation favors this interpretation but raises the question as to the source of the carbonate minerals.

In the area around San Lucas, fragments of fine-textured gray limestone are abundant on slopes bordering the arroyos. Although the source of this rock was not determined, its presence suggests that the unusually hard caliche cap around San Lucas may be a reconstituted limestone gravel.

In the Guadiana Valley the most likely source of the carbonate minerals is in the tuff of the rhyolitic series. Thinly and evenly stratified tuffaceous beds exposed in cuts along the Mazatlán highway commonly produce a weak effervescence when treated with hydrochloric acid. Moreover, the bare surfaces of tuff in foothills west of Durango themselves have veinlets and crusts of caliche in their weathered zones. Although no lentils of freshwater limestone or other visible concentrations of carbonate minerals were seen in the Tertiary volcanic series, the carbonate is possibly a minor primary constituent of such portions as may have been deposited in standing bodies of water.

Brown Paleosol

At nearly every exposure of the Guadiana formation, the caliche has been partly or entirely leached from the uppermost part. The depth of leaching ordinarily ranges between a few inches and several feet. The leached zone is occupied by a reddish-brown soil, which owes its color to quantities of ferruginous clay.

Exposures in the gravel pits at Locality 24 best show the relationships between the caliche and the brown soil (Pl. 2, fig. 3). Here the contact between the two units is irregular, veinlets and films of the reddish clay extending downward into the caliche, and corroded pinnacles of the caliche standing as residuals in the brown soil. In places the leached zone extends as much as 3 feet into the caliche.

Similar effects of leaching may be seen in exposures alongside the Mazatlán highway at Locality 3. Pebbly sand and clay near the top of the Guadiana formation are light gray because of an abundance of caliche in the form of nodules and irregular vertical columns several inches long. From the upper foot of this material the carbonate has been dissolved away, leaving a dark reddish-brown silty clay which extends downward into the calcareous substratum as numerous illuvial films and veinlets.

Some of the more permeable deposits of course gravel appear to have lost almost all their caliche, as for example at the type locality of the Guadiana formation where the secondary carbonate forms thick crusts coating bedrock at the base of the gravel. Also at Cruces (Fig. 1) the caliche is not now concentrated at the top of the gravel but is dispersed as thin lenses throughout an 18-foot section.

This iron-enriched layer, with its illuvial clay, is identified as the B horizon of a mature humid soil (*see* Carter and Pendleton, 1956, p. 489–490). The brown soil evidently records an interval of moist climate following the dry stage that produced the caliche.

Younger Basaltic Lava

Flows of fresh black basalt appear to be most extensive in the part of the Guadiana Valley southeast of the Torreón highway (*see* U. S. Air Force Aeronautical Chart and Information Service World Aeronautical Chart 521, Lake Santiaguillo sheet, 1955). The craggy surfaces of these more recent flows are unweathered, and they preserve in many places the corrugations and other original surficial features of the lava. Cactus and scrub grow in crevices over the hummocky surface, which is characterized by domical hillocks rising 20 feet or so above subcircular depressions a few tens of feet in diameter.

At Locality 29 the southern end of the younger basalt field ends at a slough bordering the cypress marsh of the Mezquital flood plain. The slough has cut its bed about 10 feet below the surface of the lava, but there is nothing here to indicate that the basalt ever extended southward to form a natural dam across the Mezquital drainage.

The rock at this locality is scoriaceous olivine basalt. A thin section of a specimen taken from a horizon a few feet below the original surface of the lava shows hyalopilitic texture, with random phenocrysts of fresh olivine and plagioclase in dark glass containing scattered microlites of feldspar. The olivine is chrysolite, and the composition of the plagioclase is near Ab_1-An_1, as determined by optical methods. Glass makes up about half the rock.

This lava is classified as younger than the Guadiana formation principally because it is

unweathered, whereas the paleosols developed upon the Guadiana indicate that this formation has been subjected to protracted weathering under both dry and humid conditions.

Pueblito Formation

Definition.—Pueblito formation is the name here given to alluvial and lacustrine deposits which in places rest unconformably on the Tertiary volcanic rocks, but more commonly overlie the Guadiana formation disconformably. The maximum exposed thickness is 28 feet, although the formation may be thicker than this in the lower parts of the Guadiana Valley. By contrast with the Guadiana formation the Pueblito is on the whole finer-textured, containing more sand, silt, and clay than gravel. The younger alluvium contains much humic material, and it lacks the conspicuous secondary accumulations of caliche so characteristic of the Guadiana. The few fossils that have been taken from the Pueblito formation belong to the Recent epoch.

The type section is in the steep south bank of the Río del Tunal at the settlement of Pueblito (Locality 2; Pl. 3, fig. 3).

Type Section of the Pueblito Formation
(Locality 2, Fig. 2)

Unit	Thickness (feet)
7. Silt, gray; compact toward base, friable toward top	1.5
6. Silt and fine sand, mixed; tiny pellets of caliche in upper part. Contains concretionary bodies like those of Unit 4	2.0
5. Silt, gray, compact; lentil of pebbles at base. Contains concretionary bodies like those of Unit 4	1.5
4. Pebble gravel grading upward into sand and silt. Contains many long cylindrical ferruginous concretionary bodies one-quarter to half an inch in diameter, standing at right angles to bedding	3.5
3. Sand, gray; grades upward into compact silt. Contains concretionary bodies like those of Unit 4	5.5
2. Sand, coarse, with lentils of granule and pebble gravel	3.0
1. Gravel, brownish; of subangular to well-rounded pebbles of volcanic rock; matrix sandy. Base not exposed	6.0
Total exposed thickness	23.0

This section represents almost the entire thickness of the formation. Rhyolitic bedrock is probably no more than a few feet below the exposed parts of Unit 1. Unit 7 forms the surface of a narrow terrace bordering the Tunal where it issues from the mountains and enters the Guadiana Valley. Exposures of the formation are continuous for about half a mile along the river at Pueblito.

Exposures.—Within the drainage of the Tunal, the Pueblito formation is exposed in the banks of nearly every stream, as well as in many of the road cuts and irrigation ditches. Exposures of 10 or more feet of section may be seen at Localities 8, 9, 10, 19, and 20. Thinner sections are exposed at Localities 5, 15, 16, 17, 22, 23, 25, and 27. Except at Locality 19, where the silt is mud-cracked at one horizon, these exposures add little to what can be gained from an examination of the type locality.

Within the drainage of the Sauceda, along the banks of the arroyo entering the reservoir at Miguel Hidalgo from the south, a gravelly phase of the Pueblito formation is exposed. At Locality 32 the formation consists of 15 feet of interbedded gravel, sand, and silt capped by 1 foot of dark-gray humic silt. Eleven feet of interbedded sand and pebble gravel may be seen along the same arroyo farther downstream at Locality 34. The coarser texture of the formation in this area possibly reflects the conglomeratic nature of source beds in Tertiary strata forming the lower parts of hills south of Miguel Hidalgo.

Arroyos trenching the western piedmont of the Sierra de la Silla have cut through the Pueblito formation and into the Guadiana formation in many places (Pl. 3, fig. 1). The Pueblito in this area is predominantly gray to brownish-gray clayey or sandy silt, locally with lentils of interbedded pebble and cobble gravel. A dark humic zone, 10–18 inches thick, caps the section.

Like the Guadiana formation, the Pueblito continues northward into the drainage of the Río de San Juan. At San Lucas, 7 feet of humic clayey silt overlies the hard caliche at the top of the Guadiana formation. At Cruces the formation consists of gray silt with interbedded sand and gravel, which fills an old channel cut 18 feet through the Guadiana formation (Pl. 3, fig. 2).

Deposits of white marl at Rancho Chapultepec (Locality 50, fig. 2) are here included with the Pueblito as a lacustrine phase of the formation. These deposits occupy shallow depressions several hundred yards in diameter that are surrounded by low ridges of the older basaltic lava. Most of the depressions were

marshy or dry during the drought of 1956, but wave-cut benches surrounding the basins at horizons about 10 feet higher than their lowest parts mark the high-water levels of these ephemeral lakes. Gullied slopes above the high-water benches lead up to the rims of the basins which are 5.5 feet above the high-water levels. The gullies are cut in white silty marl containing abundant shells of lacustrine gastropods, which are mingled with the bones of terrestrial vertebrate animals near the margins of the basins.

Relationship to younger lava flows.—At Locality 29 the top of the Pueblito formation overlaps the edge of the younger basalt flows. It follows that these parts of the formation are younger than this particular field of basalt. Obviously the age relationships may be otherwise with respect to older parts of the Pueblito, or with respect to younger basalt flows elsewhere.

Fossils.—The fossils in the Pueblito formation are all Recent. From the exposures at San Lucas the party collected an ungual phalanx, a fragment of the first phalanx, and a fragment of the ulna of modern bison. The humerus of this same animal was dug from the alluvium at Cruces. Cursory collecting from marginal parts of the lacustrine deposits at Rancho Chapultepec produced teeth and bone fragments of *Equus caballus laurentius* (Hay 1913) (Quinn 1957), numbers of small bird and frog bones, and parts of the spotted skunk, *Spilogale lucasana* Merriam (1890).

Dr. James Quinn, who made these identifications, remarked that since the spotted skunk lives in this part of Mexico today and is not found elsewhere, its bones may have been introduced into the lacustrine deposits through burrowing. He has observed, furthermore, that all the bones taken from the Pueblito formation release the "burned-bone odor" on ignition. To his knowledge the oldest bone that releases this odor is dated at around 6800 B.P.

Gastropods from the lacustrine beds include *Helisoma trivolvis lentum* (Say), *Lymnaea palustris* (Sowerby), and *Succinea grosvernori* Lea. Dr. E. P. Cheatum, who is responsible for these identifications, pointed out that one of the shells of *Lymnaea* shows distinct malleations, suggesting the buffeting action of waves.

Buried hearths and artifacts.—The humic zone which forms the upper foot of the Pueblito formation in the broad valley west of the Sierra de la Silla was an occupation level for human cultures. At a site bordering the eastern shore of the artificial lake in this valley, Dr. Kelley found numerous flakes of flint and a few stone projectile points weathering from this zone. No pottery could be found. Buried hearths, again without associated pottery, weather from this same zone at near-by sites. According to Kelley (personal communication, March 7, 1958) the culture represented in these preceramic sites is the Las Caracoles. This is a local phase of the Southwestern Desert Cultures, the western regional branch of the North American Archaic Stage.

Origin.—Much of the fine-textured sediment in the Pueblito formation may have been reworked from upper horizons of the brownish soil developed upon the Guadiana formation. This soil would have been the most accessible source of silt and clay in those areas where the Pueblito fills channels scoured into the older alluvium. Where the Guadiana formation was largely or entirely removed by erosion prior to the succeeding episode of alluviation, as in the valley to the south of Miguel Hidalgo, coarse conglomerate of the Tertiary bedrock was exposed and reworked into the Pueblito.

The abundance of humus in the alluvium argues for deposition during a stage of moist cool climate. Long cylindrical concretionary bodies that stand vertically in the beds of silt at the type locality and elsewhere suggest fillings in tubelets left by the decay of reedy plants, which were periodically buried by new increments of graded sediment. The fact that evidences of human occupation appear to be restricted to the uppermost layer of the Pueblito, an incipient humic soil formed after the valley flats had been leveled by alluviation, suggests that marshy conditions, inimical to settlement, prevailed in the lowlands while most of the Pueblito was being deposited.

If, as here suggested, the lake deposits in the older lava fields are contemporaneous with the dark alluvium of the valleys, these afford further evidence of humid climate, and suggest that the Pueblito belongs to one of the postglacial pluvial stages.

Surficial Deposits

General features.—Lower slopes of bedrock hills bordering the Guadiana Valley are for the most part covered with stony mantles that have crept downward from the croppings above. These colluvial deposits are succeeded basinward by thin blankets of silty alluvium, which in most places cover the Pueblito and Guadiana formations to depths ranging between

a few inches and several feet. Coarser alluvium in transit along channels of the larger streams forms discontinuous flood-plain deposits that rarely exceed 10 feet in thickness. Where the wind has blown without obstruction across flood plains, braided channels, or the beds of ephemeral lakes, the finer sediment is commonly reworked into dunes. All these sedimentary accumulations, with the exception of the colluvium, are demonstrably younger than the Pueblito formation.

Colluvium.—Colluvium 3–9 feet thick mantles hill-sides of andesite along the highway to Torreón between Localities 46 and 47. As exposed in the sides of a quarry at Locality 46, the colluvium consists of andesite boulders embedded in dark-gray silty clay. The base of the bouldery beds cleanly truncates prominent veinlets of caliche that extend several feet down into the bedrock. In the road cut at Locality 47 a colluvial mantle 9 feet thick is sharply divided into two parts, the lower 5 feet consisting of andesite boulders encrusted with caliche, and the remainder of dark colluvium like that at Locality 46. Caliche in the older colluvium possibly corresponds in age with the caliche developed upon the Guadiana formation, but the physical connection, if any, between either of the two superposed mantles and the alluvium of the bordering valley has not been established.

Silty deposits of the plains.—The lower reaches of the Guadiana Valley are for the most part floored with a blanket of silt covering an erosional surface developed upon the Guadiana formation in some places and upon the Pueblito in others. The silt is characteristically gray, friable and massive; locally it is dark gray and humic, and in some places it is interbedded with sand and gravel. Near Durango City this material may be seen overlying the Guadiana formation at Localities 3, 4, 6, 7, 13, 24, 25, 26, 31, and 39 (Fig. 2; Pl. 2, fig. 3). Also along the Torreón highway, northeast of Locality 47, the silt in most places covers the Guadiana to depths of a few inches. At Localities 5, 8, 10, 16, 17, 20, and 32 the silty bed rests upon the Pueblito formation. The relief along the surface of erosion amounts to no more than a few inches at most exposures. Along the western piedmont of the Sierra de la Silla, however, the silty deposits and interbedded sand and gravel locally fill channels 3 or 4 feet deep entrenched in older alluvium (Pl. 3, fig. 1).

The silt is thus younger than the Pueblito formation. It is older than many of the present arroyos, some of which are probably re-excavated along channelways formerly filled with the silty deposits. At many places the silt contains pottery and even relics of Spanish and Mexican occupation. J. C. Kelley (personal communication, March 7, 1958) believes that the pottery represents the full range of a Chalchihuites Culture occupancy near Locality 13, and perhaps that of the subsequent Loma San Gabriel Culture as well.

Aeolian deposits.—Silty deposits of the plains drift with the wind where disturbed by plowing or depleted of their natural grassy cover by grazing animals. The southern margin of the plain in which the city of Durango is situated is a particularly dusty area, and in the vicinity of Locality 17 the wind has excavated many shallow depressions in the silt. These end downward at a tough clayey substratum in the Pueblito formation. During wet weather the wind-scoured depressions become shallow ponds and marshes.

The settlement of Arenal, on the bank of the Mezquital, is doubtless named for the blanket of wind-blown sand that extends upward from the river over the lower slopes of hills to the east. The source of the sand is the broad channel of the river, which is widely exposed during low stages. Although parts of the sandy area are still actively being moved by the wind, most of the sand is anchored by vegetation.

Fixed dunes of light-gray dust and fine sand border the eastern and northeastern sides of ephemeral lakes around Rancho Chapultepec (Locality 50, Fig. 2). The source of these wind-blown deposits is the silty lacustrine marl of the lake beds themselves. Pottery weathers from along the contact between the aeolian and lacustrine deposits, so that the dunes cannot be very ancient, even though they are presently inactive and anchored by grass.

Flood-plain deposits.—Narrow and discontinuous flood plains stand 10 to 18 feet above the channels of the Tunal and Sauceda along their courses through the Guadiana Valley. Only the largest ephemeral streams or arroyos have flood plains, although narrow alluvial benches locally occur within the channels around the inner sides of the bends.

The flood-plain deposits consist for the most part of loose sand and gravel. At Locality 45 this material is dug from the banks of the Sauceda for use in road construction.

Correlation

Quaternary Sequence in the Valley of Mexico

The intermontane basin in which Mexico City is located contains a remarkable sequence of interrelated fluviatile, lacustrine, volcanic, and glacial deposits, which has become a standard for the continental Quaternary in this part of the world. The following table summarizes the sequence of alluvial deposits and fossil soils, as described by Bryan (1948), de Terra *et al.* (1949), Arellano (1951a; 1951b; 1953a; 1953b), and Lorenzo (*in* Mooser *et al.*, 1956).

Synopsis of Quaternary Alluvial Deposits in the Valley of Mexico

——*Present*——

Recent

8. Nochebuena formation: Alluvial, lacustrine, and marsh deposits; contains pottery and artifacts at certain horizons

——*circa* 2500 years ago——

7. Totolzingo formation: Dark-gray silt and sand; contains artifacts but no pottery

——*circa* 4500 years ago——

6. Barrilaco paleosol: Caliche along erosional disconformity

——*circa* 7000 years ago——

Pleistocene

5. Becerra formation:
   - c. *Upper member:* Interbedded sand, gravel, and silt; contains remains of bison, horse, elephant, and ancient man

   ——between 11,000 and 16,000 years ago——
   - b. *Paleosol:* Caliche bands and veinlets in zone as much as 6 feet thick, developed along erosional disconformity
   - a. *Lower member:* Interbedded clay, silt, sand, and gravel; contains bones of elephant, bison, and land turtles

4. Morales paleosol: Caliche along erosional disconformity
3. Tacubaya formation: Compact brownish alluvium
2. Caliche; unnamed

Pliocene?

1. Tarango formation: Interbedded volcanic, alluvial, and lacustrine deposits.

These units vary in thickness from place to place, but the combined thickness of Units 3 through 8 is ordinarily no more than about 32 feet (Arellano, 1953b).

Most workers agree that the caliche of the Barrilaco paleosol developed during the dry Altithermal Age, 4500–7000 years ago; and there are radiocarbon dates to support these and other figures for ages given in the synopsis (*see* Lorenzo *in* Mooser *et al.*, 1956).

There is wide disagreement, however, as to the age of the Becerra and older formations. Sears and Clisby (1955) found nothing in the pollen taken from long cores of lake sediments to indicate that any part of the section, even including the supposed equivalents of the Tarango formation, is older than the Wisconsin glacial stage. On the other hand, Arellano (1953b) refers the Tacubaya to the Kansan and Nebraskan stages of the Pleistocene.

The upper member of the Becerra formation is the only Pleistocene unit of which the fauna is known in some detail. Aside from de Terra's famous Tepexpan man and the imperial mammoths associated with him, Hibbard (1955) has described an armadillo, a glyptodon, two species of gopher, a cat, a bear, several species of horse, a camel, two species of bison, and a bovid distantly akin to the mountain goat. On the evidence of this fauna, Hibbard believes that the erosional surface between the upper and lower members of the Becerra was cut during the Sangamon interglacial stage, and that the deposition of the younger Becerra alluvium began during the late Sangamon and continued during an unspecified part of Wisconsin time. *Testudo*, a land turtle not known to range above the Sangamon, is reported by Hibbard from the lower member of the Becerra.

Tentatively, the Guadiana formation may be correlated with the lower member of the Becerra formation, on the ground that both units contain *Testudo*. The caliche in the Guadiana may thus correspond to the intraformational caliche of the Becerra. The brownish paleosol of Durango is taken to represent an interval of weathering under humid conditions, during a time of nondeposition or erosion in the Guadiana Valley that corresponded with the deposition of the upper Becerra in the Valley of Mexico. By this interpretation the Barrilaco paleosol is not present in the Guadiana Valley. Perhaps the dry Altithermal Age was here represented by the episode of arroyo-cutting that produced erosional surfaces separating the Guadiana and Pueblito formations. In any event, the Pueblito allu-

vium, with its Recent fauna and pre-ceramic archeological associations, seems to be the most likely counterpart of the Totolzingo formation. The silty surficial deposits of the plains, in turn, may be compared with the Nochebuena.

Were it not for the superposition of the brownish soil upon the caliche of the Guadiana formation, this caliche would be better correlated with the extensive Barrilaco paleosol of central Mexico. According to Arellano (1953a) the Barrilaco can be recognized in most of the large valleys on the Mexican Plateau between the parallels 16° 30′ N. and 25° 30′ N. Neither Arellano nor others who have described the caliche, however, mention an iron-rich soil developed upon it. Because of this difference, and also because soil scientists are beginning to regard red color in soils as presumptive evidence of geologic antiquity, the caliche of the Guadiana Valley is believed to be far more ancient than the Altithermal Age (*see* Carter and Pendleton, 1956, p. 504–505).

Davis Mountains of Western Texas

The sequence of alluvial formations in the Guadiana Valley closely resembles that of the Davis Mountains in the Big Bend country of Trans-Pecos Texas. The Guadiana and Pueblito formations may be correlated with the Neville and Calamity formations, respectively, of the broad valleys and pediments of the Davis Mountains. Moreover, the superficial silty beds above the Guadiana and Pueblito formations appear to correspond with the Kokernot formation of Texas (Albritton and Bryan, 1939; Bryan, 1941).

Like the Guadiana formation, the Neville contains the bones of elephants, has abundant caliche, has veinlets of brownish illuvial clay passing through the carbonate-rich zones, and has to date yielded no remains of human culture.

Like the Pueblito formation, the Calamity is set disconformably into the older alluvium, contains much humus, and is associated with buried hearths and artifacts of the Southwestern Desert Cultures. The snail *Helisoma trivolvis lentum* (Say) is common to both deposits but has not been reported from below either. To date the Calamity has yielded a far greater variety of human remains and artifacts than has the Pueblito, and its buried hearths occur throughout its thickness, which averages 12 feet and ranges up to 30 feet.

The Kokernot formation consists largely of friable sand and silt covering valley flats to average depths of a foot or so but locally filling channels as deep as 13 feet. In its relationship to the older bodies of alluvium, it is identical with the superficial silty layer on the plains around Durango City. Like these silty deposits, the Kokernot contains pottery, potsherds from its upper layers having been dated as 1100–1400 A.D.

Part of this correlation has already been anticipated by Arellano (1951a, p. 62; 1953a), who first suggested that the Becerra formation is "a more or less close equivalent" of the Neville, and who later suggested that the caliche in the Neville might correspond to the intraformational caliche of the Becerra rather than with the younger Barrilaco pedocal.

Summary and Discussion

Beginning probably at some time during the late Tertiary or early Quaternary, an extensive erosional surface in western Durango was dissected. Its remnants, preserved as broad divides between the canyons of the Sierra Madre Occidental, slant eastward toward the Mesa del Norte. Originally this surface probably drained into the Rio Grande, because most of the fish in headwaters of mountain streams now draining to the Pacific belong to the Rio Grande fauna.

In the area of the Guadiana Valley the southern headwaters of the Pleistocene Rio Grande have been captured by streams draining to the Pacific Ocean. At least two species of tropical fish have migrated from the Pacific lowland into the Guadiana Valley up the Mezquital and Tunal Rivers (Meek, 1904, p. xxxvii). The deadly scorpion of Durango is also a tropical migrant that followed the valley of the Mezquital upward from the coast of Nayarit (Rouaix, 1929, p. 27–28).

Precisely when this capture occurred remains an open question. The level of the Guadiana Valley must have been lowered by erosion several hundred feet below the upland surface of the Sierra Madre before the pediment flanking the eastern slope of the Sierra del Registro could have formed. This pediment was graded to a broad valley which may still have drained into the Rio Grande.

The canyon incised below the Registro pediment at the falls of the Mezquital may mark the initial capture of the Rio Grande headwaters. This canyon was later filled to depths approaching 200 feet by the older basaltic

lava, after which the Guadiana formation was deposited upon the basalt. If one may accept the presently recognized range of *Testudo*, the cutting of the canyon and the effusion of lava were both pre-Wisconsin events.

The Guadiana formation, a coarse-textured accumulation containing bones of large grazing animals, was probably deposited under humid conditions. On the assumption that humid intervals of the Pleistocene coincided generally with glacial stages, and by analogy with the chronicle of the southern High Plains of the United States, the Guadiana is more likely to belong to the Illinoian or some earlier glacial stage than to the Sangamon interglacial. The great concentrations of caliche in the Guadiana alluvium would more likely belong to the Sangamon, which Frye and Leonard (1957) characterize as a generally dry interval for the southern High Plains. So interpreted, the brown paleosol developed upon the caliche of the Guadiana formation may be the only lithologic record of the Wisconsin glacial stage.

The interval of erosion following the development of the brown soil may have coincided with the dry Altithermal Age of about 4500 to 7000 years ago. Subsequent filling of the gullies with the humic silt of the Pueblito formation proceeded under more humid conditions, when, if this synthesis is correct, numerous small lakes filled depressions in the older fields of lava.

It would be premature to date the more recent geologic events in advance of Dr. Kelley's report on the archeology. On the evidence of radiocarbon determinations now available, Kelley (personal communication, March 7, 1958) would place the pre-ceramic sites of the uppermost humic layers in the Pueblito alluvium somewhere within the interval 6000–500 B.C. The friable gray silt on the surface of the Guadiana Valley contains pottery related to Chalchihuites occupancy *circa* A.D. 500–1350.

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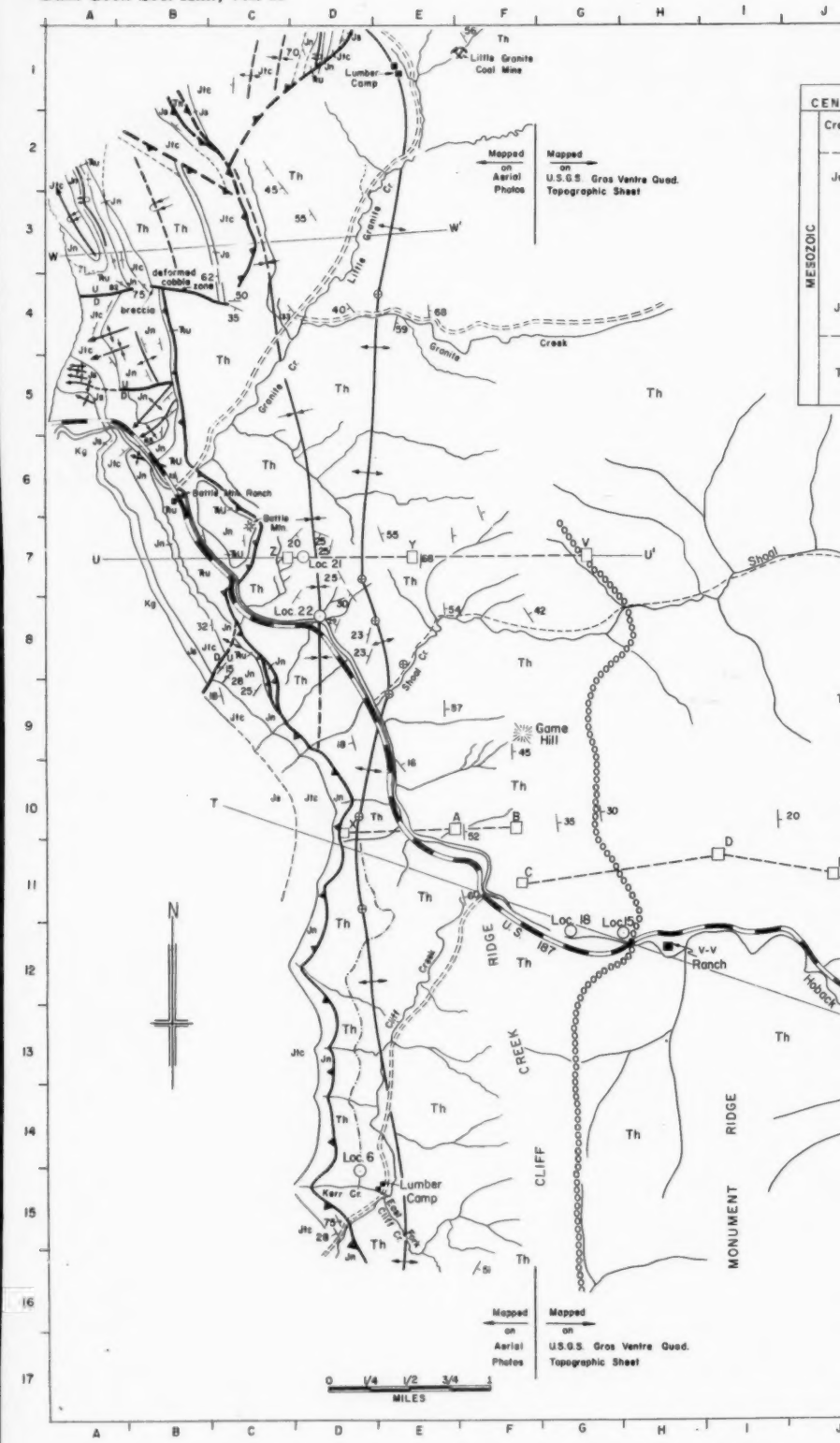
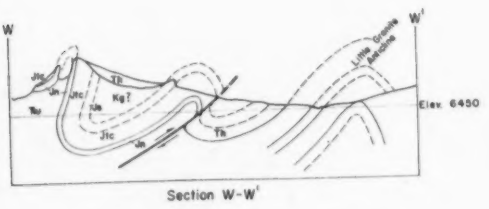
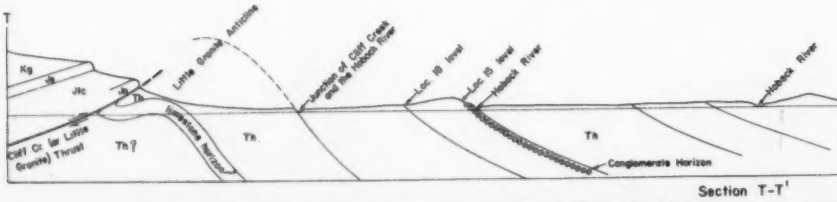






GEOLOGIC MAP AND CROSS SECTIONS OF

J K L M N O P Q R S

STRATIGRAPHIC COLUMN
(For Detailed Map Area Only)

CENOZOIC	Paleocene & Earliest Eocene	Hoback Formation	Th
MESOZOIC	Cretaceous	Early Cretaceous (& /or Latest Jurassic)	Gannett group (= Cloverly with basal part ? = Morrison)	Kg
	Jurassic	Late Jurassic	Stump Formation (="Upper Sundance")	Js
		Middle & Early Late Jurassic	Preuss Formation	Jtc
			Twin Creek Formation (="Lower Sundance")	
			Gypsum Spring Formation	
	Jurassic (or Late Triassic)	Early Jurassic	Nugget Sandstone	Jn
	Triassic	Late Triassic	Timothy Formation of Wanless, et al, 1955. (= "Lower Popo Agie" =Upper Chugwater = Ankareh, & locally including Deadman Limestone or "Upper Popo Agie")	Tru

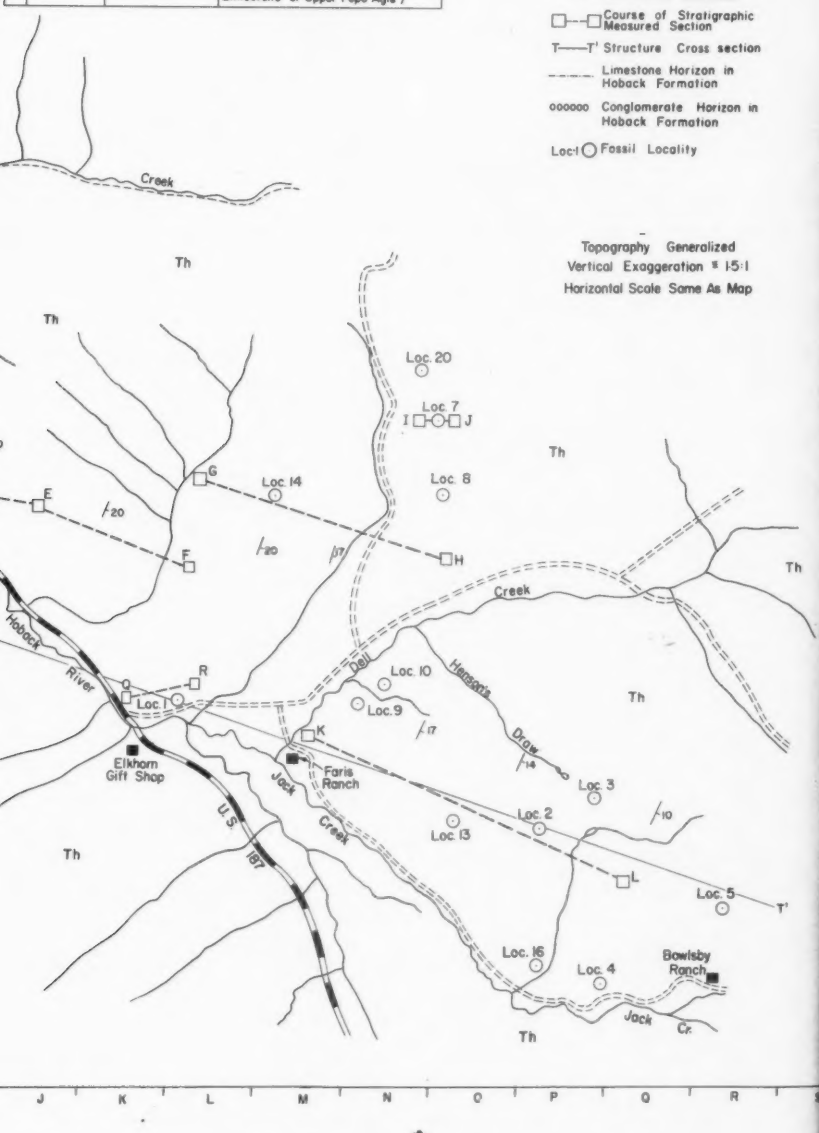
CULTURE

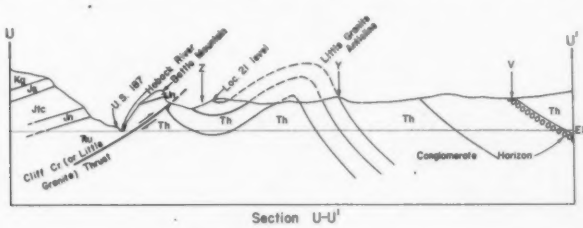
- Forest Service Trail
- Roads
- Streams

GEOLOGIC SYMBOLS

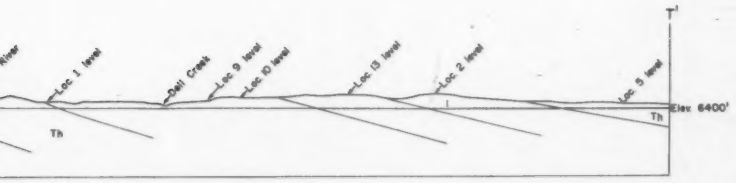
- Contact
- Fault
- Thrust Fault
- Overturned bedding
- Vertical bedding
- Horizontal bedding
- Anticline showing plunge
- Syncline showing plunge
- Overturned Anticline
- Course of Stratigraphic Measured Section
- T—T' Structure Cross section
- Limestone Horizon in Hoback Formation
- ooooo Conglomerate Horizon in Hoback Formation
- Loc.1 Fossil Locality

Topography Generalized
Vertical Exaggeration = 1.5:1
Horizontal Scale Same As Map





Section U-U'



OF PART OF HOBACK BASIN

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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EARLY CENOZOIC VERTEBRATE PALEONTOLOGY, SEDIMENTATION, AND OROGENY IN CENTRAL WESTERN WYOMING

BY JOHN A. DORR, JR.

ABSTRACT

Study of the early Cenozoic Hoback formation adds detail to the Laramide history of central western Wyoming. The recently discovered late Torrejonian, Battle Mountain vertebrate faunule, low in the formation, includes a new condylarth, *Tetraclaenodon transitus*. Additions to the intermediate Dell Creek faunule include a new insectivore, *Elpidophorus minutulus*. Additional mammals from three other localities are described. In combination with earlier studies, stratigraphic thickness, faunal ages, and faunal levels provide estimates of sedimentation rates and times of orogeny.

The Hoback formation is a thick deposit of fluviatile, palludal, and lacustrine materials that accumulated in a subsiding orogenic trough. Sedimentation was very rapid, probably beginning and accelerating in Torrejonian, culminating during late Torrejonian, then decelerating during Tiffanian, Clarkforkian, and Graybullian times prior to a late phase of orogeny. Sediment was derived locally from western, mid-Laramide highlands which began to rise in the early Torrejonian; the uplift culminated between the middle and end of the Torrejonian. Orogenic phases were relatively brief but intense. The area of deposition was much lower, forested, temperate, humid, locally swampy with some lakes, and largely inhabited by a forest-dwelling mammalian fauna. The Cliff Creek thrust which overrode the Hoback formation is internally complex, with distinct fold-fault sets trending parallel with, perpendicular to, or acute to the thrust trace. Fold axes in the overridden western part of the Hoback formation roughly parallel the thrust trace and are concealed locally beneath the thrust. Local deformation severely distorted the conglomeratic materials in the formation.

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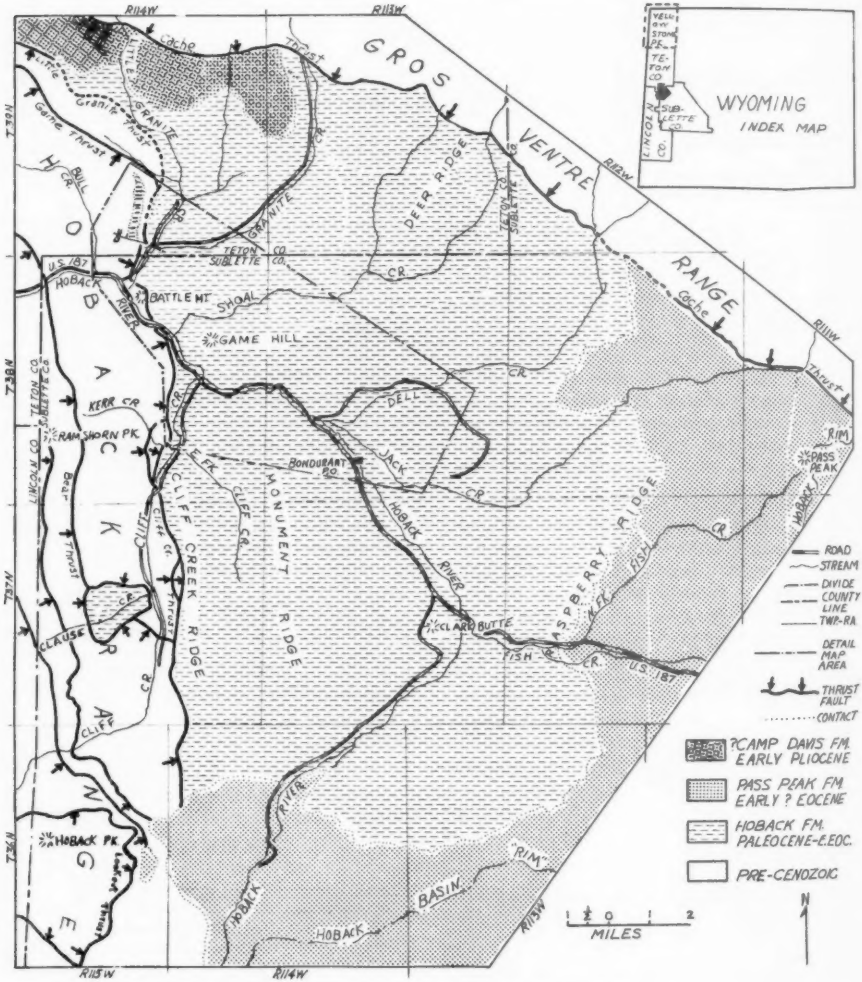

FIGURE 1.—INDEX AND GENERAL GEOLOGIC MAPS
Revised from Dorr (1952) as modified from Eardley (1944).

INTRODUCTION

General Statement

The Hoback Basin (Fig. 1) forms the northern tip of the Green River structural basin, but the two basins are physiographically distinct. Roughly triangular, the Hoback Basin occupies about 250 square miles in the northwestern corner of Sublette County and small adjoining portions of Lincoln and Teton counties, Wyoming. The Hoback and Gros Ventre ranges converge to form its northern apex. To the west is the Hoback Range, to the north and northeast the Gros Ventre Range. The basin broadens southward and is separated from the Green River Basin by a low drainage divide called The Rim. U. S. Highway 187 enters the basin from the south over a low point on the divide.

Rocks in the bordering ranges represent all periods from Precambrian through Cretaceous, possibly excepting the Silurian (Wanless *et al.* 1946; 1955, P. 1).

The basin lies approximately on an ancient

sedimentary axis that separated relatively thin foreland facies to the east in Wyoming from thick geosynclinal facies to the west in northern Utah and southern Idaho during Paleozoic and Mesozoic times (Wanless *et al.* 1955, p. 76–77). The abrupt change in thickness between those two facies greatly affected the development of Laramide structures (Dorr, 1956, p. 99, 107).

Several thick sections of Cenozoic strata, derived from Laramide and later highlands, and in some places yielding datable vertebrate fossils, were caught in the grip of Laramide and later orogenic forces in a manner that permits the dating and to some extent the analysis of Cenozoic orogenic events.

Summary of Earlier Investigations

Detailed reviews of the geology of the Hoback Basin and vicinity appear in the recent works of Dorr (1952; 1956), A. J. Eardley, L. Horberg, V. E. Nelson, and V. Church (1944, privately printed maps and cross sections), Eardley and White (1947), Eardley (1951), Horberg *et al.* (1949), Love (1956a; 1956b), and Wanless *et al.* (1955). This area was relatively quiescent throughout the Paleozoic and Mesozoic; sedimentation proceeded more or less without interruption. The stratigraphic nomenclature applied in the region to deposits of those eras is complicated by the interfingering of foreland and geosynclinal facies (Wanless *et al.*, 1955, p. 7).

Eardley (1951, p. 323 and Fig. 215) has inferred that during the latest Cretaceous and earliest Paleocene, folding and thrusting of an initial phase of the Laramide orogeny may have occurred in Idaho west of the Hoback Basin. Such an event is not clearly evident in the Cenozoic deposits of the Hoback Basin. Finer clastics in the lowest-exposed portions of the Hoback formation may possibly indicate that initial phase of orogeny, but the age of those levels is still unknown.

A second phase of orogeny (mid-Laramide of Eardley, 1951, Fig. 215) followed during the Paleocene when the Absaroka, St. John, and Darby thrusts developed in the Snake River and Wyoming ranges of eastern Idaho and westernmost Wyoming. The rising orogenic belt shed clastic debris eastward into the Hoback Basin area to form the thick Hoback formation.

Deposition of this formation continued without interruption from the Paleocene into the earliest Eocene, when the first stage of the late Laramide orogenic phase began. In the Hoback Range just west of the Hoback Basin, a series of north- to northwest-trending imbricate thrust sheets rode eastward. Chief among these, from west to east, are the Bear, Game, and Cliff Creek (known also as Jackson or Little Granite) thrusts. The Cliff Creek thrust cut and overrode the Hoback formation, so the thrusting clearly occurred after the earliest Eocene (post-Hoback formation). Shortly thereafter, a flood of dark, metaquartzitic, pressure-marked Beltian (Precambrian) cobbles and pebbles entered the area, building up the Pass Peak formation. These materials were no doubt reworked from a Paleocene deposit, the Pinyon conglomerate, to the north (Dorr, 1956, p. 106; Love, 1956b, p. 144). The metaquartzites of the Pinyon conglomerate were derived from Beltian rocks to the north (Eardley, 1951, p. 323). Reworking of these materials from the Pinyon conglomerate into the Pass Peak formation resulted from late early Eocene uplift in Teton County to the north (Love, 1956b, p. 144). The Gros Ventre Range, which now borders the Hoback Basin on the northeast, did not exist then, or it would have blocked the entry of the Pass Peak conglomerates (Eardley, 1951, Fig. 182; Dorr, 1956, p. 108). The Pass Peak formation overlapped the thrusts of the First Stage of the Late Phase of Laramide orogeny along the east side of the Hoback Range (west side of the Hoback Basin). Renewed thrusting in that range cut the Pass Peak formation. Eardley (1951, p. 323 and Fig. 215) terms this movement the Second Stage of the Late Phase of Laramide orogeny. It occurred in post-Pass Peak time. Although the age of the Pass Peak formation is not known with certainty, fossil plants suggest that it is partly late early Eocene (late Wasatchian), so this stage of orogeny must have begun after late early Eocene time. Invertebrate fossils suggest that deposition of the Pass Peak formation may have continued into the middle Eocene (Bridgerian), in which case the orogenic movement might even be post-Bridgerian. The Gros Ventre Range along the northeast margin of the basin formed much later, during the early Pliocene.

Purpose and Procedure of This Investigation

The foregoing sketchy account serves as a background for this paper. Much of the geologic history of the Hoback Basin and vicinity remained unclear after the author's first publication on the area (1952). The Hoback formation, a synorogenic deposit resulting from mid-

Laramide orogeny, promised to provide additional data bearing on the nature and dates of orogeny.

New types of fossil vertebrates were found at several known localities. Two especially important new fossil vertebrate localities (21 and 22) furnished specimens collectively called the Battle Mountain local fauna. The faunule helps date the lower part of the Hoback formation, provides new information on Paleocene faunas, and includes some new forms of mammals.

Field mapping on aerial photographs and topographic maps revealed some details of structure within the Cliff Creek thrust and in the overridden Hoback formation.

Acknowledgments

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Vertebrate Paleontology

General

The Hoback formation is the only Cenozoic deposit in the Hoback Basin that has yielded identifiable vertebrate remains. Intensive search since 1952 for fossils in the Pass Peak and Camp Davis formations has been fruitless. The author (1952, p. 72–73) listed 19 fossil localities in the Hoback formation, six of which had provided identifiable vertebrate fossils. The faunules from two of those six localities, the Dell Creek Quarry (UM-Sub-Wy Loc. 1) and UM-Sub-Wy Loc. 2, have particular stratigraphic significance. The Dell Creek local fauna comes from about two-thirds of the way up in the exposed portion of the formation and establishes the beds at that level as Tiffanian (early late Paleocene; *see* Fig. 2). The UM-Sub-Wy Loc. 2 faunule comes from near the top of the exposed portion of the formation and establishes the beds at that level as earliest Eocene (early Wasatchian-Graybullian). Since 1952 the author has collected additional specimens from several old localities and from three new ones, UM-Sub-Wy 20, 21, and 22 (Pl. 1).

The specimens from localities 21 and 22, collectively called the Battle Mountain local fauna, come from about the same stratigraphic level, and the localities are geographically close. The level is low in the exposed part of the formation. The base of the formation is probably not far below, but it is concealed beneath the Cliff Creek thrust that overrode the formation from the west.

Taxonomic descriptions of the new specimens, grouped according to locality, and a summary of the paleontologic significance of the Hoback formation faunules follow. The faunules are treated in order of age from oldest to youngest. The letters CM before a catalogue number stand for Carnegie Museum, the letters UMMP for the University of Michigan Museum of Paleontology.

Batile Mountain Local Fauna (UM-Sub-Wy Localities 21, 22)

The author discovered these fossil localities after his initial publication on the area (1952), so the specimens are described in this paper for the first time, although the fauna and one especially well-preserved condylarth from Locality 22 were mentioned in a field trip guidebook (Dorr, 1956, p. 106). The fossils from localities 21 and 22 are combined under the Battle Mountain local fauna because they apparently occur at the same stratigraphic level and are only a short distance apart geographically.

Locality 21, from which all but one of the specimens came, is an exposure of gray and brown clays on the crest of a ridge about 500 yards east-southeast of the steep east face of Battle Mountain. The U. S. Geological Survey topographic map of the Jackson Quadrangle does not show section lines in the township in which the locality occurs, but section lines from bordering townships can be projected to give a close approximation of the location

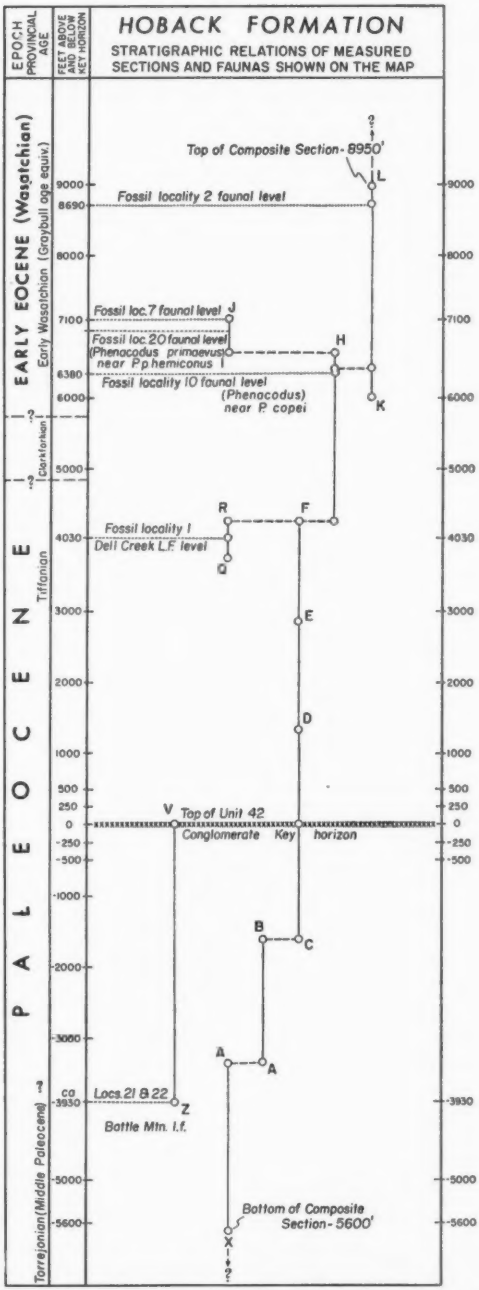

FIGURE 2.—STRATIGRAPHIC LEVELS OF SOME VERTEBRATE LOCALITIES IN THE HOBACK FORMATION

as in the NW¼ NW¼ sec. 10, T. 38 N., R. 114 W., Sublette County, Wyoming. The exposure lies east of U. S. Highway 187 across the Hoback River and is visible from the highway.

Locality 22 is south of Locality 21 and southeast of Battle Mountain in about the NW¼ SW¼ sec. 10, T.38 N., R. 114 W., Sublette County. The single specimen, a condylarth (UMMP 34762), came from the base of the upper one-third of a meander scar cut by the Hoback River. The scar is just across the river from U. S. Highway 187 at a turnout in the highway once known as the Old Thompson Dude Camp. Plate 1 shows the locations and the inferred structural and stratigraphic relationships of the two localities yielding this faunule.

Class OSTEICHTHYES

Order Indeterminate

Pharyngeal teeth of an unidentified bony fish are common. Several uncatalogued specimens are in the Carnegie Museum collection from Locality 21. The specimens are significant only because they imply a fluvial or lacustrine environment of deposition.

Class REPTILIA

Order CHELONIA

Family Indeterminate

Turtle carapace fragments are common at Locality 21. Combined with the fish teeth, mammals, and invertebrate shell fragments, they suggest fluvial deposition.

Class MAMMALIA

Order MULTITUBERCULATA

Family PTILODONTIDAE

Subfamily PTILODONTINAE

Ptilodus near **P. montanus** Douglass

Ptilodus montanus DOUGLASS, 1908, Carnegie Mus. Annals, v. 5, p. 14–15; *see* SIMPSON, 1937, U. S. Nat. Mus. Bull. 169, p. 84–94 for full synonymy

Three fragmentary specimens establish the presence of the genus *Ptilodus* in the faunule. They probably belong to different individuals of the same species. They are near *P. montanus* in the characters they show collectively, but none of the specimens is adequate in itself for such close identification.

UMMP 34763 is a fragmentary P_4 with the following features: strong, unbranched lateral ridges; no broad groove on anterior edge; high, rounded lateral profile as in *Ptilodus*; three equally spaced and separate ridges from the first serration; a basal concavity (indicating the presence of P_3). The length cannot be measured, but the width (measured as in Jepsen, 1940, p. 248) is about 1.3 mm.

UMMP 34764 is a tricuspid P^1; length 2.6 mm, width 2.1 mm.

CM 8805 is a P^3 with six cusps; length 3.2 mm, width 2.2 mm.

Order INSECTIVORA

Family LEPTICTIDAE

Cf. **Leptacodon** Matthew and Granger, Species Indeterminate

Leptacodon MATTHEW AND GRANGER, 1921, Am. Mus. Novitates no. 13, p. 2–3

UMMP 34765 is the trigonid of P_4?. The cusps are high and sharp; protoconid is higher than metaconid; paraconid-protoconid crest is relatively near the midline and removed from the external margin of the tooth. It seems certain to be the tooth of a leptictid and compares closely in structure with the various species of the genus but is too incomplete for specific identification. It is larger than *L. ladae* Simpson, *L. munusculum* Simpson, *L. tener* Matthew and Granger, or *L. packi* Jepsen. The paraconid is strong, unlike the reduced cusp in similarly sized *L. siegfriedti* Simpson.

Order PRIMATES

Family PLESIADAPIDAE

Pronothodectes matthewi Gidley

Pronothodectes matthewi GIDLEY, 1923, U. S. Nat. Mus. Proc., v. 63, p. 12

UMMP 34766 is a well-worn left M^2. The tooth is transverse and trapezoidal, with nonlobed base as broad as outer tooth margin. Paraconal and protoconal corners are extended. Anterior and posterior margins are nearly parallel and transverse to the tooth row. Protoconule and metaconule are absent. Length is 2.0 mm, width 3.2 mm. The specimen compares closely in size and construction with the type figures and description of *P. matthewi*. It is nearly identical with two specimens (35463 and 35470) of that species in the collection of the American Museum of Natural History, differing from them only in lacking the conules (which are very weak in those specimens).

Plesiadapis gidleyi (Matthew)

Nothodectes gidleyi MATTHEW, 1917, Am. Mus. Nat. History Bull., v. 37, p. 832
Plesiadapis gidleyi (Matthew) SIMPSON, 1935, Am. Mus. Novitates, no. 817, p. 3

Table 1 gives the ranges of the measurements of the Battle Mountain specimens. These may be

compared with Simpson's statistics for *P. gidleyi* from the early late Paleocene Tiffany fauna from the San Juan Basin. In all comparable measurements except M_1 length and the M_3 noted above, the ranges of the two groups of specimens either overlap broadly or coincide. In the case of M_1 length the measurements of two Battle Mountain specimens extend beyond the upper limit of the range of the Tiffany group. However, the difference in width of M_1 between the smaller of the two Battle Mountain specimens and the mean of that measurement in *P. gidleyi* from the Tiffany is only 2.25 times the standard deviation of the mean of *P. gidleyi*. Additional specimens from Battle Mountain might possibly demonstrate an overlap in ranges for that dimension, too. The two groups of specimens are alike in size and indistinguishable in structure and therefore probably belong to the same species. Such differences in size range as are shown by the present collections are of subspecific value at most.

Table 1.—Measurements (in mm) of Plesiadapis Gidleyi in the Battle Mountain Local Fauna

(UMMP 34767–34773 and CM 8806, 8807, 8823, 8828, 8830)

	P_3		P_4		M_1		M_2		M_3		M^1		M^2		M^3	
	L	W	L	W	L	W	L	W	L	W	L	W	L	W	L	W
Observed range	2.0	1.7	2.2–2.3	2.0	3.2–3.4	2.55–2.8	3.0–3.4	3.0–3.4	4.6–4.9	2.65–3.3	3.1–3.3	4.6–4.9	3.3	5.3–5.4	3.2	4.5
Number of specimens	1	1	1	1	2	2	5	5	3	3	2	2	2	3	1	1

A dozen fragmentary specimens of this genus and species are in the combined collections of Carnegie Museum and the University of Michigan, making this primate the most abundantly represented member of the faunule.

The most complete specimen, UMMP 34767, is also the smallest individual. It consists of several separated teeth, P_3 to M_3, found in close association in the matrix. In spite of its small size, however, all the dimensions except the width of M_3 fall within the ranges of individual tooth measurements given by Simpson (1935*a*, p. 6) for *Plesiadapis gidleyi*.

Order CARNIVORA

Suborder CREODONTA

Family Arctocyonidae

Thryptacodon near T. australis Simpson

Thryptacodon australis Simpson, 1935, Am. Mus. Nat. History Novitates, no. 817, p. 20–22

Three specimens of this creodont are in the collections from Battle Mountain. They are: CM 8817, a lower jaw fragment with the talonid of M_2?, talonid width 5.7 mm; CM 8811, a fragment of left M^2; and UMMP 34774, a talonid of right M_2, talonid width 5.4 mm. The teeth are closest in size to *T. australis*, being smaller than *T. antiquus* Matthew and *T. olseni* Matthew, and larger than *T. belli* Gazin. The complete cingulum around the hypocone distinguishes them from *T. belli* and *T. demari* Gazin. Comparable parts are not preserved in the type of *T. pseudarctos* Simpson. The fragmentary nature of the specimens rules out a positive identification of the species, but there is no doubt that the genus represented is *Thryptacodon*.

Order CONDYLARTHRA

Family Hyopsodontidae

Subfamily Mioclaeninae

Litaletes sp.

Litaletes Simpson, 1935, U. S. Nat. Mus. Proc., v. 83, p. 242

UMMP 34775 is a left M^1? somewhat weathered. Although a portion of the cingulum is broken away, the approximate measurements of the tooth are: length 3.5 mm, width 4.8 mm.

Allowing for weathering damage, the tooth is most like M^1 in *Litaletes disjunctus* Simpson in size and form. Yet the specimen differs from that species in that its hypocone is more distinct and relatively larger, its anterior cingulum terminates more abruptly, and its conules are not joined by thin ridges to the para- and metacones. The first of these differences makes the specimen appear slightly more progressive than *L. disjunctus*. Among the other genera of small Paleocene hyopsodontids with which it compares at all closely, the specimen is distinguished from *Litomylus* Simpson by a continuous cingulum from the hypocone to the metastylar apex; from *Litolestes* Jepsen by its marked rectangularity; and from *Haplaletes* Simpson by its lack of a bilobed inner

base. Although the specimen probably represents a new and slightly more progressive species of *Litaletes*, it is inadequate for the type of a new species.

Subfamily HYOPSODONTINAE

Haplaletes near **H. disceptatrix** Simpson

Haplaletes disceptatrix SIMPSON, 1935, U. S. Nat. Mus. Proc., v. 83, p. 244

UMMP 34776 is a right maxillary fragment with M^2 complete and the roots of P^4, M^1, and M^3. It compares most closely with *H. disceptatrix* but is smaller, more transverse, and with a sharper metastylar corner. It may represent a new species of the genus, but it is inadequate for a type specimen.

Family PHENACODONTIDAE

Gidleyina? cf. **G. wyomingensis** Gazin

Gidleyina wyomingensis GAZIN, 1956, Smithson. Misc. Coll., v. 131, no. 6, p. 42

The following specimens form the basis for the identification: UMMP 34777, LM_1 in jaw fragment; UMMP 34778, LM_3; UMMP 34779 RM_2? in jaw fragment; UMMP 34780, heel of RM_1 and RM_2 in jaw fragment; CM 8808, RM_1; CM 8815, RM_2 in jaw fragment; and CM 8820, RM_2?. Identification was difficult and remains uncertain, because only lower molars are in the collection; the diagnostic premolars were not found. Diagnoses of the genera *Gidleyina*, *Ectocion*, *Tetraclaenodon*, and *Phenacodus* do not adequately distinguish the molar teeth of those groups. Each genus is composed of several species having molar teeth with size and structural variations which may overlap not only those of other species in the genus but even those of species in other genera. In many places these condylarths are the most abundantly represented members of middle and late Paleocene and early Eocene faunas, the wealth of material is now unwieldy and urgently calls for statistical treatment. Such treatment is not attempted here, but some notes on the problem of distinguishing these genera follow.

Of all the species in the several genera mentioned, the specimens compare most closely in size and form with *G. wyomingensis*. They are especially alike in their low, weak paraconids and weakly basined trigonids, thus differing from *G. superior* Simpson. The cingulum in the Battle Mountain specimens differs no more from the type of *G. wyomingensis* than does the cingulum within the population including the type of *G. wyomingensis* described by Gazin (1956a) from the Bison Basin, Wyoming. The author agrees with Gazin's suggestion (1956a, p. 43) that *G. superior*, *G. silberlingi* Simpson, and *G. montanaensis* Simpson may be synonymous. Thus it is less difficult to assign the Battle Mountain specimens to a species than to a genus, a situation commonly encountered in dealing with Paleocene mammals.

The species represented probably belongs to the genus *Gidleyina*. Although size alone is not a valid generic character, the species of *Phenacodus* and *Tetraclaenodon* do tend to be larger. The species of *Ectocion* are close in size to those of *Gidleyina*. Structurally, however, there are differences in the molar teeth that in combination may be used to distinguish between *Gidleyina* and the other genera. The lower molars of *Gidleyina* have an enclosed, basined trigonid and a strongly twinned entoconid. Examination of specimens in the U. S. National Museum and the American Museum of Natural History showed that incipiently twinned entoconids and enclosed trigonid basins occur on some specimens of the several species of *Phenacodus* and *Tetraclaenodon*, although the accessory cuspule on the entoconid is in all cases weak. The species referred to *Ectocion* have trigonids that are not basined and not enclosed by the parastylid crest. A very weakly twinned molar entoconid appeared on M_3 of only one specimen (AMNH 16061) of *E. ralstonensis*. Only one specimen of *E. osbornianus* (AMNH 16047) showed such twinning. The premolar entoconids of *Ectocion* may be twinned.

The ranges of molar dimensions in the Battle Mountain specimens are:

M_1 (two specimens)	L—6.7–6.9 mm W—5.2–5.3 mm
M_2 (three specimens)	L—6.8–7.0 mm W—5.2–5.7 mm
M_3 (one specimen)	L—7.6 mm W—4.9 mm

PLATE 2.—NEW SPECIES OF FOSSIL VERTEBRATES, AND DEFORMED COBBLES FROM THE HOBACK FORMATION

FIGURES 1–4.—*Tetraclaenodon transitus* n. sp. Holotype. Late Torrejonian (late middle Paleocene) Battle Mountain local fauna. UMMP 34762. All figures *ca.* × 1. (1) Right P^3 to M^3, occlusal view. (2) Same as Figure 1, labial view. (3) Right lower jaw with P_4 to M_3, labial view. (4) Same as 3, occlusal view.

FIGURES 5–7.—*Elpidophorus minutulus* n. sp. Holotype. Tiffanian (early late Paleocene) Dell Creek local fauna. UMMP 34758. Left lower jaw with P_2 to M_2. P_3 and P_4 crushed below normal occlusal level. All figures *ca.* × 5. (5) Lingual view. (6) Labial view. (7) Occlusal view.

FIGURES 8–11.—Deformed cobbles from a faulted and folded conglomerate horizon in the Hoback formation. Figures 8, 9, one specimen; 10, 11 another specimen. All figures *ca.* × ½.

NEW SPECIES OF FOSSIL VERTEBRATES AND DEFORMED COBBLES FROM THE HOBACK FORMATION

Depth of mandible below lingual alveolar lip of M_2 trigonid (two specimens)—9.7–12.2 mm

Genus **Tetraclaenodon** Scott, 1892

Tetraclaenodon transitus n. sp.

(Plate 2, figures 1–4)

HOLOTYPE: UMMP 34762. Badly crushed, incomplete skull with right P^3 to M^3 and left P^2 to P^3; incomplete right mandibular ramus with P_4 to M_2 and alveoli of P_3 and M_3. Collected by J. A. Dorr, Jr., August 17, 1954.

HORIZON AND TYPE LOCALITY: Late middle Paleocene (late Torrejonian), Battle Mountain local fauna, Hoback formation, UM-Sub-Wy Loc. 22 (Pl. 1). Meander scar along Hoback River southeast of Battle Mountain, in about the NW ¼ SW ¼ sec. 10, T.38N, R. 114W., Sublette County, Wyoming.

DIAGNOSIS: Much larger than *Tetraclaenodon puercensis* (Cope), *T. pliciferus* (Cope), *T. subquadratus* (Cope), and *T. symbolicus* Gidley (by Simpson from Ms.), and thus the largest species yet referred to the genus. Molars distinctly transverse, as is the tendency in *Tetraclaenodon*, but mesostyle present on M^1–M^3 as in *Phenacodus* (and in some specimens of *Tetraclaenodon*). Trivial name alludes to the species' intermediate position between *Tetraclaenodon* and *Phenacodus*. Measurements (in mm) of the type specimen are:

P^2	L—9.3	P_4	L—11.8
	W—*ca.* 10.3		W—9.0
P^3	L—*ca.* 9.4	M_1	L—12.0
	W—*ca.* 11.5		W—10.5
P^4	L—9.4	M_2	L—12.1
	W—12.3		W—10.8
M^1	L—11.2	P_4–M_2	35.8
	W—14.6 plus (broken)		
M^2	L—11.2		
	W—15.1		
M^3	L—10.0		
	W—13.4		
M^1–M^3	—33.4		

DISCUSSION: Matthew (1937, p. 188) extensively revised the diagnosis of *Tetraclaenodon* to include consideration of the whole skeleton as well as the teeth. However, fossilized skeletal parts other than teeth are rare, and only the teeth are well preserved in UMMP 34762. In this specimen, the characters Matthew recognized as distinguishing the cheek teeth of *Tetraclaenodon* from those of the later genus *Phenacodus* are merged in such a fashion as to suggest that the new species described here is intermediate, although closer to the older *Tetraclaenodon*. The present writer compared the specimen with the types and many other specimens of species of both *Tetraclaenodon* and *Phenacodus* in the collections of the U. S. National Museum and the American Museum of Natural History. In terms of Matthew's diagnoses, the specimen resembles *Phenacodus* in its possession of a distinct (although relatively weak) mesostyle on the upper molars. Furthermore it is distinctly larger than any described species of *Tetraclaenodon* and clearly falls within the size range of the population of Clarkforkian phenacodonts, probably *Phenacodus primaevus*, analyzed by Simpson (1937b, p. 17–19), although the proportions of the teeth differ. It should be noted that a weak mesostyle occurs in some specimens of *Tetraclaenodon puercensis*, so that characteristic must be used with caution.

On the other hand, the specimen is allied to *Tetraclaenodon* by the following characters: P^3–P^4 relatively short and broad, with small metacone subordinate to and confluent with the paracone; upper molars short, broad, and decidedly transverse, especially M^2; the lower molars that are preserved (M_1 and M_2) have a vestigial paraconid on the anterior flank of the metaconid. Aside from their difference in size, *Tetraclaenodon transitus* and *T. puercensis* resemble one another more closely than either resembles *Phenacodus primaevus*, especially in that in both the upper molars are transverse, the talonid of P_4 is narrow, the lower molars are not as inflated, and the transverse protoconid-metaconid crest is nearly perpendicular to the anteroposterior midline of the lower molars.

Simpson (1935a, p. 22–23) noted that Gidley had postulated the existence in the Tiffany (early late Paleocene) of a genus intermediate between *Tetraclaenodon* and *Phenacodus*. Simpson concluded that the Tiffany specimens were not intermediate but were generically identical with *Phenacodus*. The present writer believes that *T. transitus* represents an intermediate form such as Gidley postulated, but that the transition from *Tetraclaenodon* to *Phenacodus* was accomplished without the interpolation of a third genus.

UMMP 34781 includes badly worn P^2 to P^3 of a specimen that seems referable to the new species. It comes from Locality 21 (not 22), tending to confirm the conclusion, already supported by stratigraphic and structural evidence, that the two localities are not only close to one another geographically but are also at practically the same stratigraphic level.

Tetraclaenodon Species Indeterminate

UMMP 34782 is an isolated right M_2. It is clearly smaller than the Clarkforkian and Graybullian *Phenacodus* and smaller than *T. transitus* but is larger than *T. puercensis*. It compares closely in size with *P. grangeri* from the Tiffany fauna but differs from that species in the retention of a paraconid. It seems to be *Tetraclaenodon* but represents a second species of that genus in the Battle Mountain local fauna. The specimen is too incomplete for assignment to a species.

Additions To The Dell Creek Local Fauna (UM-Sub-Wy Locality 1)

Class REPTILIA

Three fragmentary lizard jaws (CM 8878, 8882, and 8888) were added to the collection from this

quarry. A similar jaw was previously reported. These new specimens are mentioned so that some future student of such reptiles may know of their existence. The author is unable to determine their affinities.

Class MAMMALIA

Order MULTITUBERCULATA

Family PTILODONTIDAE

Genera and Species Indeterminate

The ptilodontid material collected from the Dell Creek Quarry since 1952 is of poor quality. None of the specimens can be identified with certainty, although most of them probably belong to *Ptilodus*, a genus already reported from the quarry. However, two specimens are of special interest.

CM 8885 is an upper premolar, probably P^1, that is unique, so far as can be determined from the literature, in that it is bicuspid rather than tricuspid. Possibly it is an abnormal tooth. If not, it is from an undescribed form. The specimen is too incomplete to be made the type of a new genus and species.

CM 8883 is a small P^3 with five cusps. It is 2.2 mm long, 1.3 mm wide. It is too small to belong to any described species of *Ptilodus*. *Ectypodus*, the only other genus known from the quarry, has fewer cusps on that tooth. It seems, then, that there is still another ptilodontid in the fauna. Further collecting may provide more complete material.

Order INSECTIVORA

Family DELTATHERIDIIDAE

cf. **Gelastops parcus** Simpson

Gelastops parcus SIMPSON, 1935, U. S. Nat. Mus. Proc., v. 83, p. 227

The only other known occurrence of this species is at the type locality, where it is so poorly represented that little is known about the nature of its variation. An LM_3, CM 8802, is the sole, and admittedly weak, evidence for the presence of this genus and species in the Dell Creek local fauna. However, direct comparison shows that it is identical in size and construction with the same tooth in the type specimen of the species, USNM 6148, from the middle Paleocene Silberling Quarry, Crazy Mountain Field, Montana.

Family LEPTICTIDAE

Leptacodon near L. **ladae** Simpson

Leptacodon ladae SIMPSON, 1935, U. S. Nat. Mus. Proc., v. 83, p. 228

This author previously recorded *Leptacodon packi* Jepsen from the Dell Creek Quarry on the basis of a single specimen (UMMP 27235). A second leptictid jaw fragment, UMMP 34757, with worn left P_4 to M_2 is now in the collection. This new specimen is difficult to place taxonomically. P_4 is 1.7 mm long, only slightly larger than in UMMP 27235. Length from P_4 to M_2 is 4.6 mm, which is the same as in the type of *Leptacodon ladae* from the middle Paleocene Gidley Quarry, Crazy Mountain Field, Montana. The specimen is structurally close to both *L. packi* and *L. ladae* but closer to the latter. Because of its fragmentary nature the specimen cannot be definitely assigned to *L. ladae*, but it does suggest that a second and slightly larger leptictid was present in the fauna.

Leptictids, Genera and Species Indeterminate

Several new specimens, too incomplete for certain assignment, nevertheless indicate a diversity in insectivores of the leptictid type that is not apparent from the list of identifiable genera and species alone.

CM 8804, a left M^2, is clearly of insectivore type. It is closest in structure to those of the leptictids *Diacodon* Cope and *Myrmecoboides* Gidley; yet, except for the more deeply hooked parastylar apex and much smaller size, it also resembles molars in pantolestids such as *Aphronorus* Simpson and *Bessoecelor* Simpson. It is about the right size to go with lower teeth belonging to *Leptacodon*, but no associated upper and lower teeth of that genus are known. The specimen is probably a leptictid, possibly either *Leptacodon* or *Diacodon*.

CM 8803 is either P_2 or P_3. It is of the size and construction one would expect to find in the diminutive *Leptacodon munusculum* Simpson from the Gidley Quarry, Crazy Mountain Field, Montana. Unfortunately, the premolars are not preserved in the type and only known specimen of that species, so direct comparison is impossible.

Family MIXODECTIDAE

Genus **Elpidophorus** Simpson

Elpidophorus SIMPSON, 1927, Am. Mus. Nat. History Novitates, no. 268, p. 5

Elpidophorus minutulus n. sp.

(Plate 2, figures 5–7)

HOLOTYPE: UMMP 34758. Left mandibular ramus with P_2–M_2. Teeth unworn, but jaw crushed so that P_3 and P_4 are depressed below normal level, and talonids are obscured. Collected by J. A. Dorr, Jr., Univ. of Michigan, August, 1954.

HORIZON AND TYPE LOCALITY: Upper Paleocene (Tiffanian), Dell Creek Quarry, Hoback formation, UM-Sub-Wy Loc. 1 (Pl. 1). Along Dell Creek Road near U.S. Highway 187. NW¼ sec. 29, T.38N., R. 113 W., Sublette County, Wyoming.

DIAGNOSIS: Smaller than *E. minor* Simpson 1937, *E. patratus* Simpson 1936, or *E. elegans* Simpson 1927. P_2 and P_3 with well-developed anterior accessory cusps. Otherwise structure is as in other species of the genus. Trivial name, *minutulus*, refers to very small size of the species. Measurements (in mm) of the holotype are:

P_2—length 1.2, width 0.5
P_3—length *ca.* 1.2, width 0.8
P_4—length 1.35, width not measureable
M_1—length 1.5, width 1.3
M_2—length 1.7, width 1.3

DISCUSSION: Simpson (1937a, p. 130–131), discussing the possible relationship of mixodectids to plagiomenids, noted the following characteristics: (1) a tendency toward elevation of internal over external cusps on the lower molars found in both families but pronounced in *Elpidophorus*; (2) displacement of hypoconulid toward entoconid found in mixodectids, (3) furrowing and cusp proliferation on cheek teeth of plagiomenids. In those three respects in particular, as well as in most other aspects mentioned by Simpson, the specimen described is a mixodectid, although the hypoconulid displacement is very slight. This author takes exception to one item in Simpson's analysis, however. He describes a tendency, supposedly found in both families, but pronounced in mixodectids as follows

"lines joining protoconid to metaconid and hypoconid to entoconid parallel and anterointernal-posteroexternal."

If the long axis of the cheek tooth row is taken as a base of reference, the present author is unable, after study of figures and specimens of members of both families, to find such a condition occurring consistently in mixodectids. In *Eudaemonema cuspidata* Simpson the lines are not parallel and are anteroexternal-posterointernal. In *Elpidophorus minor* Simpson the lines are parallel but anteroexternal-posterointernal. In *E. patratus* Simpson they are parallel but anterointernal-posteroexternal. In *Microsyops elegans* Matthew they are parallel but anteroexternal-posterointernal. In *Cynodontomys latidens* the lines are neither parallel nor alike in orientation.

The specimen described here clearly belongs to the genus *Elpidophorus*, showing as it does the following characters of the genus listed by Simpson (1937a, Table 28):

(1) Paraconid on P_4 relatively large and projecting;

(2) Talonid on M_2 (concealed on M_1) well basined and tricuspid;

(3) Internal lower molar cusps very definitely higher than external cusps;

(4) Metaconid very slightly anterior to or at least in line with protoconid on molars.

Order PRIMATES

Family PLESIADAPIDAE

Plesiadapis farisi Dorr

Plesiadapis farisi DORR, 1952, Geol. Soc. America Bull., v. 63, p. 79

Several additional plesiadapid specimens are now known. They are CM 8873, RM^1–RM^3; CM 8881, P^4; CM 8884, RM^3; CM 8889, LM_2 to LM_3. There is no apparent reason why these cannot belong to the large plesiadapid previously described as *P. farisi* in this fauna. However, the dimensions of the new specimens in at least half the cases are closer to *P. fodinatus* Jepsen and less close to *P. gidleyi* (Matthew). The structural distinctions, other than size, between *P. farisi* and *P. fodinatus* still seem valid, but those two species are evidently more closely related than this author originally proposed (Dorr, 1952, p. 80, 81).

Gazin (1956a, p. 24; 1956b, p. 7) has noted that the posterolabial slope of the protoconid on the lower molars of certain species of *Plesiadapis* tends to be short. He suggested that these species may represent a group of subgeneric rank (unnamed). *P. farisi* would belong to the group of species including *P. gidleyi*, *fodinatus*, *cookei*, *rubeyi*, and *dubius*, in which the slope is short. The specimens of those species of *Plesiadapis* examined by this author appeared to support Gazin's suggestion to the extent that the entoconid slope is either long or short, without intergradation. However, this author could not find any other character that was consistent with the grouping this one character seems to support.

Order CARNIVORA

Suborder FISSIPEDIA

Family MIACIDAE

Didymictis sp.

Didymictis COPE, 1875, U. S. Geog. Survey, W. 100th Meridian (Wheeler), p. 5, 11
Viverravus WORTMAN (AND MATTHEW), 1899, Am. Mus. Nat. History Bull., v. 12, p. 136 (*non Viverravus* Marsh, 1872)
Didymictis MATTHEW, 1915, Am. Mus. Nat. Hist., Bull., v. 34, p. 17–19

UMMP 34759 is a right M_1 that surely belongs to *Didymictis* but is from a much smaller form than *D. dellensis* Dorr from the same quarry. It is smaller than *D. protenus* (Cope) and its subspecies, *D. haydenianus* Cope or *D. altidens* Cope, but is much larger than *D. microlestes* Simpson or *D. tenuis* Simpson. The tooth very closely resembles an unnumbered specimen in the U. S. National Museum, identified by Gazin (1956b, p. 9–10) as *Didymictis*? sp. The measurements of M_1 in the two specimens

are: UMMP 34759—length 6.0 mm, width of trigonid 4.0 mm; USNM specimen—length *ca.* 6.3 mm, width *ca.* 3.3 mm. Although the length-width ratios of the teeth are different, the teeth are close to one another in size. Neither specimen is adequate to be made a type specimen of a new species, but it is evident that in both the Almy and Hoback formations there is an intermediate-sized late Paleocene *Didymictis*. It was to be expected that the genus would eventually prove to have had greater diversity than the collections showed at this level, and future discoveries will no doubt provide material which more adequately represents this small species.

Fossils From UM-Sub-Wy Locality 10

This locality (Fig. 2; Pl. 1) was mentioned in the author's 1952 publication (Table 1, p. 72) but at that time had not yielded vertebrates. It is in a shaly horizon about half a mile northeast of the William T. Faris Ranch, on the southeast side of Dell Creek, in the NW¼ SE¼ sec. 21, T.38N., R. 113 W., Sublette County, Wyoming. In addition to pelecypods, the following fossil vertebrates have been found there:

Class REPTILIA

Order CROCODILIA

Crocodilian

An indeterminate form represented only by skull fragments

Class MAMMALIA

Order CONDYLARTHRA

Family PHENACODONTIDAE

Phenacodus copei? Granger

Phenacodus copei GRANGER, 1915, Am. Mus. Nat. History Bull., v. 34, art. 10, pt. 3, p. 344

CM 8871 consists of a right mandibular ramus with M_2–M_3, a left M^3, 2 lower premolars, and several tooth fragments. Compared with the phenacodontid materials in the U. S. National Museum and the American Museum of Natural History, this specimen is identical in form and closest in size to the large form of *P. copei* from the late early Eocene (Lysite). It is larger than specimens of the smaller and older form from the earliest Eocene (Graybull) that are referred to the same species. The author erroneously called this specimen *Hyracotherium* in a preliminary identification (Dorr, 1956, p. 101, 106).

Fossils From UM-Sub-Wy Locality 20

The author discovered this new locality in June, 1952, after his 1952 publication on the area appeared. The horizon is in the top of Unit 1 of measured section 4, Traverse I-J (Dorr, 1952, Pls. 1, 3; this paper, Fig. 2 and Pl. 1), NE¼ sec. 16, T.38N., R. 113 W., Sublette County, Wyoming. The sole specimen came from the top of a shale unit just below the second massive sandstone nose east of the road over Dell Creek-Shoal Creek Divide. The locality is about 153 feet below and about 300 yards northwest of Locality 7. The specimen strengthens the conclusion that the uintathere remains found at slightly higher Locality 7 (Dorr, 1952, p. 89; 1958) are early Eocene.

Class MAMMALIA

Order CONDYLARTHRA

Family PHENACODONTIDAE

Phenacodus privaevus hemiconus? (Cope)

Phenacodus hemiconus COPE, 1882, Am. Philos. Soc. Proc., p. 179

P. primaevus hemiconus (Cope) GRANGER, 1915, Am. Mus. Nat. History Bull., v. 34, art. 10, pt. 3, p. 339

CM 8813 includes a right mandibular ramus with P_4 to M_3 and two upper molars. Comparison with the type, topotype, and other specimens in the American Museum of Natural History indicates that this specimen is closest to *P. p. hemiconus*, especially in size and in the inflated nature of P_4 and M_1. But as Simpson (1937b, p. 17–19) has shown, positive diagnosis of the several subspecies of *P. primaevus* Cope from the Graybull Eocene is in many cases difficult or impossible, so the subspecific assignment is questioned, although there is little doubt that the specimen belongs to *P. primaevus*.

Fossils From UM-Sub-Wy Locality 7

This locality is in the NE¼, sec. 16, T.38N., R. 113 W., Sublette County, Wyoming (Fig. 2; Pl. 1). It is distinguished by a new form of uintathere recently (1958) described by the author as *Prouintatherium hobackensis*. Earlier (Dorr, 1952, p. 89–90) this animal was identified by the author as *Probathyopsis successor* Jepsen. The conclusion that the uintathere is no younger than earliest Eocene is supported now by the fragmentary remains of several other smaller mammals described below from the same quarry. These specimens do not rule out the possibility of a latest Paleocene (Clarkforkian) age for the beds at this level, but an earliest Eocene age is most probable, considering the nature of the fossils already mentioned from slightly lower levels at localities 10 and 20.

Class REPTILIA

Turtle and Crocodile, Indeterminate

Turtle carapace fragments and crocodilian teeth are common at this locality, but the author is unable to identify these remains more specifically.

Class MAMMALIA

Order MULTITUBERCULATA

Family PTILODONTIDAE

Subfamily PTILODONTINAE

Genus and Species Indeterminate

UMMP 34760 is a small fragment of P_4 too incomplete for generic assignment but certainly a ptilodontine multituberculate. This specimen is significant because so far, no multituberculates are known to have existed after earliest Eocene (early Graybullian) time.

Order CONDYLARTHRA

Family HYOPSODONTIDAE

cf. **Haplomylus speirianus** (Cope)

Hyopsodus speiranus COPE, 1880, Am. Naturalist, v. 14, p. 908
Haplomylus speirianus (Cope) MATTHEW, 1915, Am. Mus. Nat. History Bull., v. 34, art. 9, pt. 2, p. 313–314

UMMP 34761 is a small P_4, certainly of a hyopsodontid, comparing very closely with *H. speirianus*. Dimensions of the tooth are: length 3.2 mm, width 1.9 mm. As in the case of the ptilodontid material described above, this species is known from the latest Paleocene (Clarkforkian) and earliest Eocene (lower and middle Graybullian). The specimen is insufficient for more certain identification.

Family PHENACODONTIDAE

Genus and Species Indeterminate

An uncatalogued LM^3 in the Carnegie Museum Collection from this locality indicates a phenacodontid condylarth. Further collecting at the site may result in the discovery of more complete and diagnostic remains of this animal, which is larger than *Hyopsodus*, *Haplomylus*, or the small forms of *Phenacodus*.

Review of Hoback Formation Vertebrate Faunules

Faunal facies.—Table 2 lists the fossil vertebrates presently known from various levels in the Hoback formation. The faunal localities are arranged across the top of the table from lowest (oldest) at the left to highest (youngest) at the right. Many of the identifications are expressed with uncertainty. In most cases this is because the specimens are fragmentary. The author attempted to identify and include in the list all available vertebrate materials, not because they are all of great taxonomic significance, but because they contribute to the age and ecologic aspects of the faunas. In turn, those aspects are important because of their bearing on rates and environments of deposition of the Hoback formation, as will be seen in a later section of this paper.

Comparison of the two principal faunas, those from the Battle Mountain and Dell Creek localities, shows that the Dell Creek fauna is larger in number of genera and species. A count of the specimens in the collection substantiates this; however, the author discovered the Dell Creek locality first and has done more collecting there. A count of specimens also shows that there are relatively few small-sized individual animals in the Battle Mountain collection in contrast to the Dell Creek Quarry collection. This difference is probably the result of incomplete and nonrandom sampling rather than a reflection of different faunal facies, because collecting conditions were different. The Battle Mountain specimens were surface finds, and under those conditions the smaller mammal remains are soon destroyed by weathering; the Dell Creek specimens were quarried and removed from a tough, unweathered limestone matrix. Fossils of small mammals do occur occasionally at Battle Mountain, and Table 2 does show that for the most part some representatives of each order and most families occur in both faunules. The only notable exception to this is the abundance of phenacodonts at Battle Mountain and their absence in the Dell Creek Quarry. The two faunules represent similar faunal facies, that is, a forest fauna with some arboreal elements. This conclusion is strengthened by the presence of a coal seam near quarry level at Dell Creek (Dorr, 1956, p. 106).

Fossil mammal remains from all other localities except Locality 7 were surface finds, so it is not surprising that they are most commonly the remains of relatively large animals. The faunules from these localities are too small to be reliable indices to environments of deposition.

Faunal ages.—The collection from Battle Mountain is the only one in which the age needs more than summary discussion; the rest of the faunules have been analyzed in other papers (Dorr, 1952; 1956).

Table 3 presents data for the age analysis of the Battle Mountain local fauna. At first glance it may appear that categories not crossing the time boundary distribute themselves nearly evenly on each side of the Torre-

TABLE 2.—VERTEBRATE FAUNULES OF THE HOBACK FORMATION

* Indicates that associated invertebrates and/or plants are relatively abundant and well preserved: (Q) quarry, (S) surface collection.

Fauna	*UM-Sub-Wy Locality 6 (Q)	Battle Mountain local fauna, UM-Sub-Wy Localities 21, 22 (S)	*Dell Creek local fauna UM-Sub-Wy Locality 1 (Q)	*UM-Sub-Wy Locality 10 (S)	UM-Sub-Wy Locality 20 (S)	UM-Sub-Wy Locality 7 (Q)	UM-Sub-Wy Locality 16 (S)	*UM-Sub-Wy Locality 2 (S)	*UM-Sub-Wy Locality 4 (S)
Class OSTEICHTHYES									
Order Semionotoidea									
Family Lepidosteidae									
Genus *Lepidosteus* sp. indet.	×	..	..	..	..	..	..	..	×
Order indeterminate	..	×	..	..	..	..	..	..	..
Class REPTILIA									
Order Chelonia									
turtle, family, gen. and sp. indet.	..	×	..	..	..	×	..	..	×
Order Crocodilia									
Family Alligatoridae									
Allognathosuchus sp.	..	..	×	..	..	..	..	..	..
crocodilian, family indet.	..	..	..	×	..	×	..	..	×
Order Squamata									
Suborder Lacertilia									
Family indet.									
Haplodontosaurus excedens?	..	..	×	..	..	..	..	..	..
lacertilian, gen. and sp. indet.	..	..	×	..	..	..	..	..	..
Class MAMMALIA									
Order Multituberculata									
Family Ptilodontidae									
Subfamily Ptilodontinae									
Ptilodus nr. *P. montanus*	..	×	..	..	..	..	..	..	..
Ptilodus? fractus	..	..	×	..	..	..	..	..	..
Ectypodus powelli	..	..	×	..	..	..	..	..	..
ptilodontine, gen. and sp. indet.	..	..	..	..	..	×	..	..	..
Order Insectivora									
Family Deltatheridiidae									
cf. *Gelastops parcus*	..	..	×	..	..	..	..	..	..
Family Leptictidae									
Leptacodon packi	..	..	×	..	..	..	..	..	..
Leptacodon nr. *L. ladae*	..	..	×	..	..	..	..	..	..
cf. *Leptacodon* sp. indet.	..	×	..	..	..	..	..	..	..
leptictids, gen. and sp. indet.	..	..	×	..	..	..	..	..	..
Family Mixodectidae									
Elpidophorus minutulus n. sp.	..	..	×	..	..	..	..	..	..
Order Primates									
Family Plesiadapidae									
Pronothodectes matthewi	..	×	..	..	..	..	..	..	..
Plesiadapis gidleyi	..	×	..	..	..	..	..	..	..
Plesiadapis farisi	..	..	×	..	..	..	..	..	..
plesiadapid, gen. and sp. indet.	..	..	×	..	..	..	..	..	..
Family Adapidae									
Pelycodus ralstoni?	..	..	..	..	..	..	..	×	..
Family Carpolestidae									
Carpodaptes hobackensis	..	..	×	..	..	..	..	..	..
Primates, family indet.	..	..	×	..	..	..	..	..	..
Order Carnivora									
Suborder Creodonta									
Family Arctocyonidae									
Thryptacodon australis	..	..	×	..	..	..	..	..	..
Thryptacodon nr. *T. australis*	..	×	..	..	..	..	..	..	..
Suborder Fissipedia									
Family Miacidae									
Didymictis dellensis	..	..	×	..	..	..	..	..	..
Didymictis sp.	..	..	×	..	..	..	..	..	..

TABLE 2.—*Concluded*

Fauna	*UM-Sub-Wy Locality 6 (Q)	Battle Mountain local fauna, UM-Sub-Wy Localities 21, 22 (S)	*Dell Creek local fauna UM-Sub-Wy Locality 1 (Q)	*UM-Sub-Wy Locality 10 (S)	UM-Sub-Wy Locality 20 (S)	UM-Sub-Wy Locality 7 (Q)	UM-Sub-Wy Locality 16 (S)	*UM-Sub-Wy Locality 2 (S)	*UM-Sub-Wy Locality 4 (S)
Class MAMMALIA (*continued*)									
Order Condylarthra									
Family Hyopsodontidae									
Subfamily Mioclaeninae									
Litaletes sp.	..	×	..	..	..	..	..	..	..
Subfamily Hyopsodontinae									
Haplaletes nr. *H. disceptatrix*	..	×	..	..	..	..	..	..	..
Haplaletes diminutivus	..	..	×	..	..	..	..	..	..
cf. *Haplomylus speirianus*	..	..	..	..	..	×	..	..	..
Hyopsodus sp. indet.	..	..	..	..	..	..	..	×	..
Family Phenacodontidae									
Tetraclaenodon transitus n. sp.	..	×	..	..	..	..	..	..	..
Tetraclaenodon sp. indet.	..	×	..	..	..	..	..	..	..
Gidleyina? cf. *G. wyomingensis*	..	×	..	..	..	..	..	..	..
Phenacodus copei?	..	..	..	×	..	..	..	..	..
Phenacodus primaevus hemiconus?	..	..	..	..	×	..	..	..	..
phenacodont, gen. and sp. indet.	..	..	..	..	..	×	..	..	..
Order Tillodontia									
Family Tillotheriidae									
Esthonyx sp. indet.	..	..	..	..	..	..	..	×	×
Order Pantodonta									
Family Coryphodontidae									
Coryphodon sp.	..	..	..	..	..	×	×	..	..
Order Dinocerata									
Family Uintatheriidae									
Prouintatherium hobackensis	..	..	..	..	..	×	..	..	..
Order Perissodactyla									
Family Equidae									
Hyracotherium sp. indet.	..	..	..	..	..	..	..	×	..

jonian (middle Paleocene)-Tiffanian (early late Paleocene) boundary. However, the numerical comparisons at the bottom of the table tip the scales slightly but clearly in favor of a Torrejonian age. The typically Torrejonian *Tetraclaenodon* was counted twice; there are two different species in the faunule, and the age aspect of any fauna is the sum of all the separate occurrences of each genus rather than the simple sum of determinable genera. The Torrejonian aspect seems clear even if the questionable parts of ranges are counted (figures outside parentheses, Table 3). There is good reason, however, to consider the count (figures inside parentheses), excluding questionable parts of ranges, because the questionable extensions are mostly in favor of the Tiffanian and appear only for one of three reasons, (1) because fragmentary specimens of *Tetraclaenodon* have very questionably been identified from the very earliest Tiffanian, or (2) because the range of *Gidleyina* is very uncertain, possibly extending into the Torrejonian-Tiffanian transition or even back into the Torrejonian (Simpson, 1937a, Table 3 and p. 48–50), or (3) because occurrences in the Bison Basin local fauna, which Gazin (1956a) tentatively concluded was early Tiffanian but suggested might be Torrejonian or transitional, are included. The figures within parentheses more strongly favor a Torrejonian age. This author concludes that the Battle Mountain local fauna is most probably late Torrejonian but possibly lies near the boundary between Torrejonian and Tiffanian.

The age of the Dell Creek local fauna is probably middle Tiffanian (Dorr, 1952, p. 88) but may be slightly older or younger.

This author has not been able to determine

the exact ages of levels at localities 10 and 20, but they are neither earlier than Tiffanian nor later than early Eocene.

northwestern margin of the Hoback Basin was continued primarily to determine the stratigraphic and structural relationships of the

TABLE 3.—RANGE CHART FOR AGE ANALYSIS OF BATTLE MOUNTAIN LOCAL FAUNA

		Torrejonian (Mid. Paleocene)			Late Paleocene						Eocene Wasatchian (Early)		
					Tiffanian (Early)			Clarkforkian (Late)					
Ptilodus nr. *P. montanus*	Genus	×	×	×	×	?	..	..	..	..	..	..	..
	Species	×	×	×	?	..	..	..	..	..	..	..	..
cf. *Leptacodon*	Genus	×	×	×	×	×	×	..	..	..	..	..	..
Pronothodectes matthewi	Genus	×	×	×	..	..	..	..	..	..	..	..	..
	Species	×	×	×	..	..	..	..	..	..	..	..	..
Plesiadapis gidleyi	Genus	..	..	..	×	×	×	×	×	×	..	..	..
	Species	..	..	..	×	×	×	..	..	..	..	..	..
Thryptacodon nr. *T. australis*	Genus	..	..	..	×	×	×	×	×	×	×	..	..
	Species	..	..	..	×	×	×	..	..	..	..	..	..
Litaletes sp.	Genus	×	×	×	..	..	..	..	..	..	..	..	..
Hapaletes nr. *H. disceptatrix*	Genus	×	×	×	?	..	..	..	..	..	..	..	..
	Species	×	×	×	..	..	..	..	..	..	..	..	..
Tetraclaenodon transitus n. sp.	Genus	×	×	×	?	..	..	..	..	..	..	..	..
Tetraclaenodon sp.	Genus	×	×	×	?	..	..	..	..	..	..	..	..
Gidleyina? cf. *G. wyomingensis*	Genus	..	..	?	?	..	..	..	..	..	..	..	..
	Species	..	..	..	?	..	..	..	..	..	..	..	..

Forms occurring in times shown at right	Torrejonian (or earlier)		Tiffanian (or later)	
Representing Genera	7 + 1?	(7)	4 + 4?	(4)
Representing Species	3	(3)	2 + 2?	(2)
Forms formerly thought restricted to	**Torrejonian**		**Tiffanian**	
Representing Genera	2 + 3?	(5)	2 + 1?	(2)
Representing Species	2 + 1?	(3)	2 + 1?	(3)

The Locality 7 level is probably earliest Eocene (early Graybullian) (Dorr, 1958).

The Locality 2 level is probably also earliest Eocene (Graybullian), but its exact place in the Graybullian is uncertain.

STRUCTURAL GEOLOGY

General

There have been many studies of the geologic structure of central western Wyoming including the Hoback Basin area. A brief summary of the structure appears at the beginning of this paper.

Mapping of structural details along the Battle Mountain local fauna; however structural details discovered within and below the easternmost thrust sheet (Cliff Creek thrust) along the western margin of the basin deserve brief mention in their own right. The area in question falls entirely within the western one-third of that shown on Plate 1. This portion was mapped entirely in the field on aerial photographs.

Folding Beneath the Cliff Creek Overthrust

The Hoback formation is thrown into an anticline along most of the western side of the

basin. The axis of this fold extends from E-1 to E-16 on Plate 1 but continues north and south beyond the map borders. The steep flank faces eastward (*see* cross sections). No doubt this fold lay largely beneath the Cliff Creek thrust prior to erosion of that thrust sheet, and the west limb is still partially concealed (D-10 to E-16). The existence of this fold has long been known; the structure was drilled twice (unsuccessfully) for oil. The first well was located in area E-2 on the map, along Little Granite Creek. The second was drilled to about 8000 feet on Sandy Marshall Creek about 7 miles south of the southern border of the map but failed to pass completely through the Hoback formation. Love (1956b, p. 146) described a fold near the head of Little Granite Creek just north of the detailed map area, calling it the Little Granite anticline. By implication, and with justification, he suggested that the Little Granite anticline might be the northward extension of the fold described above. Love stated

"The age of at least this part of the fold is presumed to be early Pliocene."

Support for his conclusion comes from his interpretation that folding resulted when the beds involved were caught between opposing late Tertiary thrusts. However, the Hoback Basin broadens rapidly to the south, so that the only thrusts closely associated with the fold there are those that moved eastward during the late Laramide orogeny. From this, and from the fact that the fold is steep on the east, the present author concludes that this is a Laramide compressional fold throughout most of its extent, although it may have been increased or extended northward by later Tertiary (post- early Pliocene) forces.

West of the anticline lies a shallow syncline. This can be traced with a fair degree of certainty from C-3 through D-9, where it passes beneath the Cliff Creek thrust. It does not reappear. Sears (1956) independently plotted this fold on a photogeologic map.

Portions of Laramide folds in the overridden Hoback formation along the western margin of the Hoback Basin lie beneath the Cliff Creek thrust. Conceivably similar folds may lie concealed beneath the same overthrust sheet. Along much of the western margin of the basin, the present thickness of the overthrust plate increases westward toward the crest of the Hoback Range, but in one area, along the west side of Cliff Creek, it is thin, and at Clause Creek has been breached by erosion, so that the Hoback formation is revealed through a fenster (Fig. 1).

Structures Within the Cliff Creek Overthrust

Small folds and faults in the sedimentary rocks of the Cliff Creek overthrust plate do not cross the thrust trace into the overridden Hoback formation; they seem to be restricted to the thrust plate.The structures fall into the following classes (*see* Pl. 1 for locations; *see also* Fig. 3):

I. Folds within the overthrust plate
   (a) Axes roughly parallel with the thrust trace and roughly normal to the direction of thrusting. Tightly folded; overturned in direction of thrust movement.
      (1) Drag folds at thrust trace. (*See* southeast corner of A-3 and cross section W-W′.)
      (2) Folds behind the thrust trace. (*See* W. ½ of A-3, SW ¼ of B-4 to NW ¼ of B-5 and cross section W-W′.)
   (b) Axes at acute angles to thrust trace (and to thrust direction). Plunging westward. Two sets, one trending roughly northwestward, the other northeastward. (*See* areas A-5, B-5, and B-6.)
   (c) Axes roughly perpendicular to thrust trace. Plunging westward. Only two folds seem to fit this questionable category. (*See* area A-4.)

II. Faults within the overthrust plate, striking either northwestward or northeastward at angle to thrust trace and seemingly related to orientation of folds in category I(b) above. (*See* A-4 to C-4 and B-5 to C-8.)

Clearly portions of the Cliff Creek thrust plate are structurally complex. The diagram of structural trends in the Battle Mountain area (Fig. 3) relates these trends to the average direction of the thrust trace. A line was drawn between the two most widely separated points on the thrust trace (*see* Fig. 1) to determine the average trace direction. The direction of thrust movement may have been nearly perpendicular to the average direction of the trace. The rose diagram of trends on Figure 3 suggests that former stresses within the thrust plate could be analyzed. The folds parallel with the thrust trace no doubt resulted from compression parallel to the direction of thrusting. The folds and faults at an angle to the thrust trace might be related to stresses such as shear, or the combined effect of lateral compression and forward

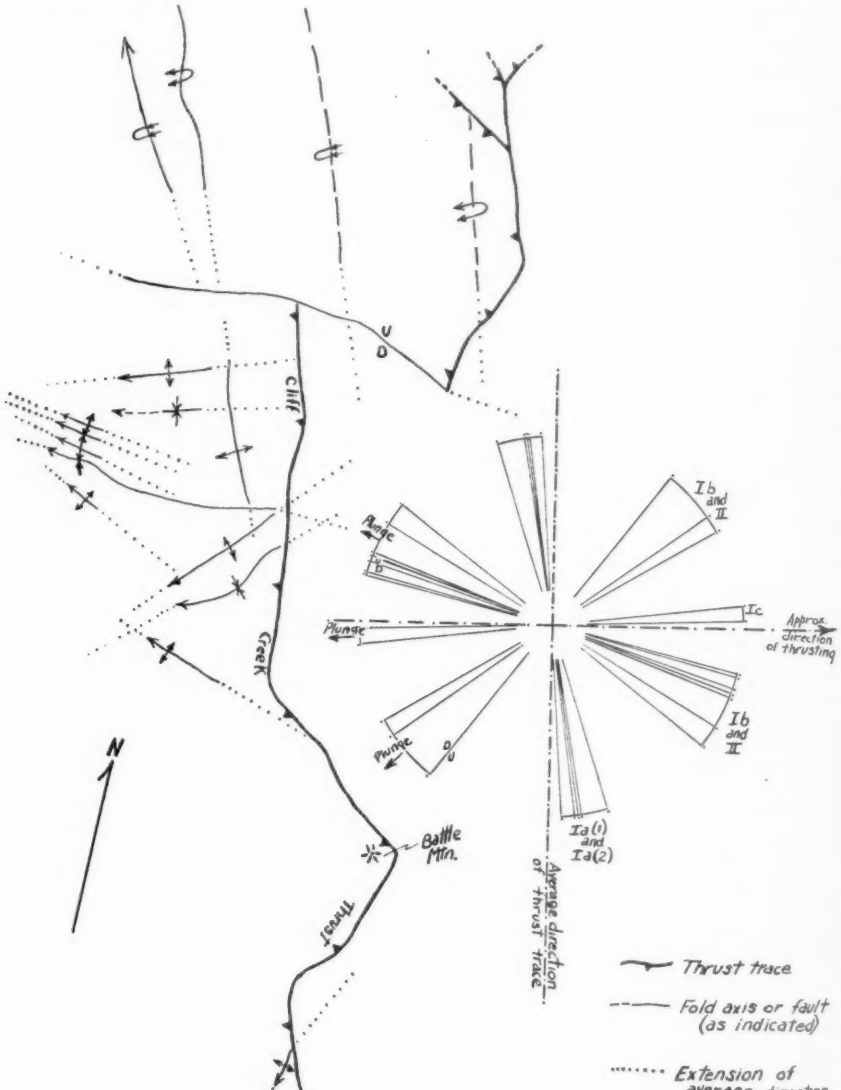

FIGURE 3.—DIAGRAM OF STRUCTURAL TRENDS IN THE BATTLE MOUNTAIN AREA

motion within the plate. Preliminary attempts at such analysis by application of the strain ellipsoid concept convinced the author that the situation is not simple. Too many essential factors are unknown, for example, the actual direction of thrust movement and its components along the thrust plane. Also, the Battle Mountain area lies close to the point where the northern portions of several eastward thrusts including the Cliff Creek thrust begin to swing around toward the northwest. This may have resulted from impingement upon the northwest-trending Precambrian crystalline axis beneath the Wind River and Gros Ventre ranges. The effects of such impingement would further complicate the analysis. At this stage the author prefers to merely describe the situation and defer the analysis for further study.

The tear fault in A-4 to C-4 is of special interest because it involved a Tertiary conglomerate. This small patch of conglomerate, labeled Th for Tertiary-Hoback formation, is preserved in a syncline within the Cliff Creek thrust on the north side of the cross fault. Evidence that the conglomerate was affected by faulting includes (1) termination of conglomerate at the fault, and (2) a concentration of deformed cobbles and pebbles at the fault (Pl. 2, figs. 8-11). There are several Tertiary conglomerate deposits in the Hoback Basin. The author tentatively identified this particular deposit as the early Pliocene Camp Davis formation and suggested that faulting occurred in the Late Tertiary (Dorr, 1956, p. 107). Further studies, including the pebble counts discussed in the following section on orogeny and sedimentation now convince the writer that the conglomerate is instead a part of the Hoback formation preserved in a syncline and carried along in the thrust. If this interpretation is correct, both the thrust and cross fault are still clearly post-Hoback (post- early Eocene), but the cross fault is not necessarily late Tertiary in age. More probably it formed at the time of thrusting.

The fault in C-8 is related to an underlying fold. This relationship is exposed in the west wall of Hoback Canyon along U. S. Highway 187 just south of Battle Mountain. The fold appears at road level. Upward in the section, above the fold and near the top of the canyon wall, the fault overlies the southwestward extension of the fold axis. A horizontal component of movement along the fault brought the southeast side over the northwest side, repeating the Jurassic Nugget sandstone in vertical section. The movement was scissorlike, with the displacement decreasing southwestward toward the pivot point. It appears that the stress which caused folding at depth within the thrust sheet was relieved by faulting above.

Orogeny, Sedimentation, and the Hoback Formation

General

The Hoback formation is a synorogenic deposit that accumulated in a downwarped basin within the Laramide orogenic belt. It owes its great thickness (about 15,000 feet) to erosion in ancient western highlands; the growth of these highlands must, in turn, have been related to the growth of folds and thrusts of some phase of the Laramide orogeny. The formation accumulated during the Laramide orogeny, because it was involved in folding and thrusting of later stages of that orogeny. Theoretically it may be desirable to distinguish between orogeny and mountain building; in the study of Laramide events this is difficult for several reasons. First, Laramide phases and smaller subdivisions involve some time intervals too small to be measured by relatively inexact paleontologic methods. Second, if any Laramide mountain building was not simultaneous with orogeny, it still occurred within a dominantly orogenic period. And third, there is good reason to believe that mountain building is not all of the postorogenic, epeirogenic type but may also be the contemporaneous result of the same orogenic forces that cause folding and thrusting. In the last case, a date for mountain building would also be an approximate date for orogeny. From this it may follow that the paleontologic dates already discussed for various levels in the Hoback formation can also be used as dates for orogeny and mountain building if the deposition of the Hoback formation can be related to those events. It is clear, of course, that basin subsidence must have accompanied mountain building and orogeny to make possible the accumulation of the thick basin sediments.

Sedimentation Rates and Orogeny

Eardley (1951, p. 287, 323–325, and Fig. 215) recognized three phases of the Laramide orogeny in the Hoback Basin area and the thrust belt of central western Wyoming and adjacent portions of Idaho. He inferred an Early Phase (latest Cretaceous and early Paleocene). No recognized deposits of this

phase are present in the Hoback Basin. In the Middle Phase (Paleocene) he tentatively included Darby, Absaroka, and some of the Bannock thrusting. Deposition of the Hoback formation and possibly of related Paleocene deposits in Jackson Hole to the north resulted from erosion of the orogenic highlands formed in this phase. A Late Phase (Eocene) is divided into two stages. In the First Stage, renewed thrusting from the west (Bear, Game, and Cliff Creek thrusts) cut the Hoback formation. The Pass Peak formation then overlapped the Cliff Creek thrust. Eardley (1951, p. 325) implied that the Pass Peak formation was derived from western orogenic highlands that rose during the First Stage of the Late Phase of Laramide orogeny. This still seems to be a logical source for the western portion of the formation, because there, in the upper Hoback River area, the formation is finer-grained, lacks reworked Pinyon-type boulders, and contains materials that may have been derived from late Paleozoic and Mesozoic rocks. However, in the eastern portion of the formation, especially in the type locality around Pass Peak, coarse Pinyon-type boulders compose much of the deposit. As Eardley no doubt recognized, these materials were eroded and redeposited from the Pinyon conglomerate of Paleocene age as a result of uplift to the north and east in Teton County. This uplift followed (or may, in part, have coincided with) the late Laramide First Stage thrusting from the west (Dorr, 1952, 1956; Love, 1956b). Thus the timing of Laramide events was conceived to be:

(1) *Early Phase*—latest Cretaceous or early Paleocene. Times of beginning, duration, and ending uncertain.

(2) *Middle Phase*—mid? Paleocene. Times of beginning, duration, and ending uncertain.

(3) *Late Phase*

First Stage—Post-Hoback formation and pre-Pass Peak formation, that is, post-Graybullian (post-earliest Eocene) and pre-Bridgerian? (pre-middle Eocene?). *Second Stage*—Post-Pass Peak formation, that is, post-early Eocene.

The dating of the Late Phase and its stages is well established (Dorr, 1952). No additional evidence bearing on the postulated Early Phase is available. However, with the help of data described in the paleontologic section of this paper, the time limits of the Middle Phase of orogeny in this region can be determined.

Figure 4 integrates paleontologic and stratigraphic data from the Hoback formation providing a means for estimating relative sedimentation rates and their bearing on orogeny. Stratigraphic thicknesses are based on measurements by the author (Dorr, 1952, and unpublished work). Because the base and top of the Hoback formation are concealed, measurements were made upward and downward from a conglomerate key horizon. Gross lithologic aspects of successive portions of the formation are plotted on the same scale as thickness. Thickness and lithologic aspect are the only quantitative data on the chart. The figure provides no means of estimating absolute time in years. The nature of exposure of the total section, measured along an 8-mile traverse, varied widely. Therefore it seemed best to express the gross lithologic aspect very generally in terms of the percentage of strongly outcropping sandstone and conglomerate to the total thickness of selected portions of the section. Thick, well-cemented sandstones are well exposed at the surface in almost all places in the area. The section below level *D* (Fig. 4) is well exposed; coal, limestone, and fine clastic material are abundant below level *A*. The section above level *D* is poorly exposed; coal, limestone, and fine clastics, seen only in scattered exposures, may be far more common than is apparent in surface outcrop. For this reason no percentages are given for those materials.

Relative time spans are from nearly identical tables found in Simpson (1947) and Wood *et al.* (1941). Their subdivisions of the Tertiary are the latest and most authoritative published. Although the nomenclature for subdivisions of the North American continental Paleocene and Eocene is nearly standard, the relative time spans are based upon subjective paleontologic judgments. However, the sequence of subdivisions is correct, and by any common standard of measurement of geologic time the Paleocene and Eocene epochs are only parts of the relatively short Tertiary Period, and the subdivisions of those epochs represent very short segments of geologic time. The complexity of mammalian structures and the rapidity with which those structures changed make it possible to recognize such brief time segments paleontologically.

The three principal vertebrate faunas from the Hoback formation are plotted on Figure 4 at their respective stratigraphic levels. Probably no two paleontologists would pinpoint any fauna at the same time; this author plots each fauna as a lens with extent in time, not to imply that there is an actual internal difference in age in any one of the faunas, but to express the

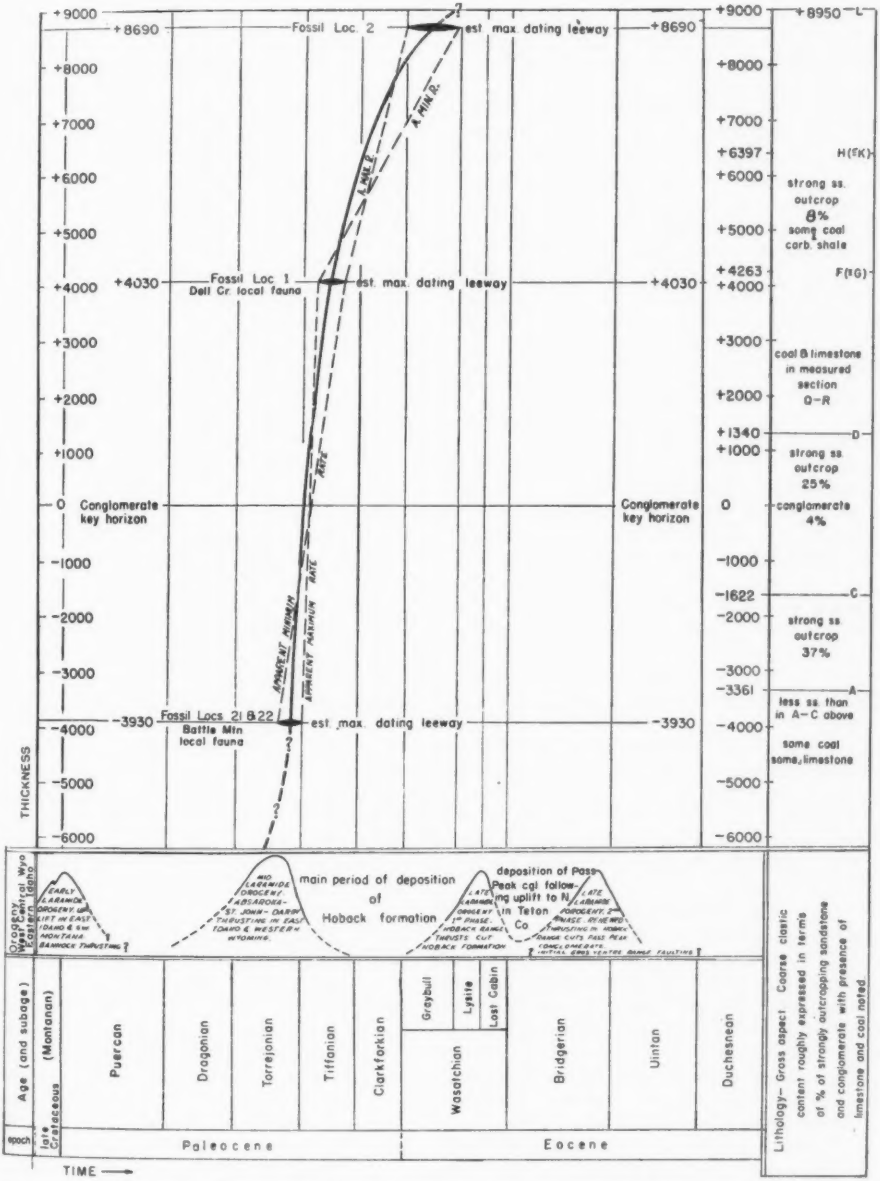

FIGURE 4.—DEPOSITION RATES, OROGENY, AND THE HOBACK FORMATION

limits of allowable dating leeway. The end points of a lens represent the oldest and youngest ages this author believes any student would assign to the fauna; the thickest point is, in the author's judgment, the most probable age.

Connecting the oldest allowable age point of a lower fauna with the youngest allowable age point for the fauna next above gives an apparent "minimum rate" for sedimentary deposition. Conversely, connecting the youngest allowable

age point of a lower fauna with the oldest age point of the fauna above gives an apparent "maximum rate". The minimum rate of the lower segment with the maximum rate of the upper segment expresses *minimum* change in sedimentation rate. Conversely, the lower maximum rate connected with the upper minimum rate expresses the maximum change in sedimentation rate.

Even the minimum change combination shows a perceptible alteration in rate of deposition of the Hoback formation. Abrupt breaks in deposition rates are, of course, artifacts stemming from the fact that there are only three faunas involved in the plotting. Had the fortunes of fossil collecting provided a middle fauna at a different level than that of the Dell Creek Quarry, the exact position of the misleadingly abrupt break in rate would be slightly different, but the over-all picture much the same. There is no *a priori* reason to believe that sedimentation followed a fluid curve in its cycle, but such a curve best serves as a generalization in the absence of any good evidence to the contrary. The upper part of the curve passes well away from the straight-rate lines. This anomaly would probably disappear if an intermediate Clarkforkian fauna were found and plotted between the two upper faunas.

The Hoback formation is about 15,000 feet thick. Figure 4 shows that most of that great thickness, 12,620 feet of sediment between the lower and upper faunas, accumulated at a very rapid rate. No doubt this was related to rapid erosion in source areas immediately to the west and that, perhaps in turn, to a short, intense orogenic spasm, the Middle Phase of the Laramide orogeny.

The curve suggests that the sedimentation rate decelerated through time. Because the Hoback formation is cut by a thrust of the First Stage of the Late Phase of Laramide orogeny, an upper limit for uninterrupted deposition is established. If the beginning of the Middle Phase of the Laramide is also the time of beginning of deposition of the Hoback formation, then a date for the base of the formation would be of great significance. The age of the base of the formation cannot yet be determined paleontologically, because the lower part is upturned and covered by the above-mentioned thrust, but the gross lithologic aspect of the formation can be used to advantage. The part of the Hoback formation between levels *A* and *D* that accumulated most rapidly is also the part wherein the highest percentage of strongly outcropping sandstone and conglomerate occur, and wherein there are no prominent coal or limestone horizons. Above level *D*, where the deposition rate seems to decrease, coal, limestone, and a lower percentage of sandstone (with more fine clastic in substitution) appear in the section. This change in lithology is correlated with a decelerating deposition rate resulting from a lowered erosion rate in neighboring orogenic highlands near the close and after the Middle Phase of the Laramide. Coal, limestone, and finer clastics also occur below level *A*. If this also reflects a slower deposition rate for the lower part of the formation, then it may indicate a lower erosion rate in the western orogenic highlands just before or in the early part of the Middle Phase of the Laramide. The time of beginning of that orogenic phase may be roughly determined by extrapolating the curve downward with a slope comparable to that of its upper portion. The slope of the curve cannot be radically flattened downward on the basis of any present evidence. If this writer's interpretation is correct, the time of acceleration in sedimentation rate of the Hoback formation cannot lie much farther back than the point in time of the lowest (Battle Mountain) fauna.

The relationship between sedimentation and orogeny is shown diagrammatically across the bottom of Figure 4. Certainly the orogenic phases were not catastrophically brief events, nor were the interorogenic times completely anorogenic. Nevertheless, the orogenic phases must have been relatively brief, closely spaced events to occur within such short allowable spans of geologic time. In summary the Laramide events are:

(1) Early Phase of orogeny; latest Cretaceous or early Paleocene?, postulated by Eardley. No evidence from Hoback Basin area. Uplift in eastern Idaho and southwestern Montana. Bannock thrusting?

(2) Middle Phase of orogeny; middle Paleocene (largely Torrejonian). St. John and Darby thrusting and related folding in eastern Idaho and western Wyoming. Orogenic highlands resulted. Orogeny culminated sometime between middle Torrejonian and beginning of Tiffanian.

(3) Deposition of Hoback formation, possibly beginning in early Torrejonian or before but reaching maximum rate at, or shortly after, culmination of Middle Phase of orogeny. Maximum accumulation during late Torrejonian and early Tiffanian. Deposition continuing

but at a decelerating rate during remainder of Tiffanian, Clarkforkian, and Graybullian times.

(4) Late Phase of orogeny—First stage; Late early Eocene. Renewed eastward thrusting in Hoback Range. Thrusts cut Hoback formation. Most active orogeny during Lysite and Lost Cabin (late early Eocene) time.

(5) Uplift to north and east in Teton County (Jackson Hole and Gros Ventre Range area) with resulting erosion of Paleocene Pinyon conglomerate material from that area and its redeposition as the Pass Peak formation in the Hoback Basin; Pass Peak formation overlapping Late Phase (First Stage) thrusts. Deposition of Pass Peak, in part during mid Eocene Bridgerian time, may have begun in latest early Eocene (Lostcabinian) time. Events 4 and 5 may have been partially contemporaneous, and some of the Pass Peak formation material may have been carried eastward from orogenic highlands formed in event 4.

(6) Late Phase of orogeny—Second Stage; late middle Eocene? (post-Pass Peak). Culminating in late Bridgerian; possibly continuing into Uintanian.

Environment of Deposition

The environment of deposition of the Hoback formation has been discussed (Dorr, 1952, p. 88–89; 1956, p. 105). In summary, in the lower two-thirds of the formation, the frequency of lithologic variation, drab coloration, crossbedding, common absence of lamination of shales, lensing character of the bedding, local limestone lenses, "dirty" sandstones and conglomerates, together suggest fluviatile-palludal-lacustrine deposition. Probably the materials were dumped by poorly graded streams, charged with sediment from the bordering highlands, upon valley flood plains in a subsiding basin of low relief. Reginald Moore (1955, unpub. M.S. thesis, Univ. of Michigan), in a recent study of the paleoecology of the molluscan fauna from the Dell Creek Quarry (Locality 1) concluded that deposition occurred at an altitude of about 1000 feet or at least much lower than the present altitude of the region, which is about 6700 feet. The pyritic limestones, with their abundant mollusc and fish remains, the black shales, and coals, probably accumulated in short-lived sloughs, oxbow lakes, and swamps. The coal and other plant remains suggest a poorly drained, forested area with a humid, temperate climate. The uniformly drab rock colors, pyritic zones, and the mammal faunas of the forest type with primates at localities 1, 21, and 22 support this inference.

In the upper, early Eocene, part of the Hoback formation, color variegation of the typical "Wasatch" type, and large mammals such as *Coryphodon* and uintatheres suggest that a change may have occurred in later times or in certain places to a savannah-like environment with glades, plains, and forested stream courses.

Source of Sediment

The materials of the Hoback formation could all have been locally derived. The quartz grains in the sandstones, although commonly showing secondary crystalline overgrowths, still possess original rounding to an extent suggesting that they were reworked from older sandstones such as the Jurassic Nugget sandstone and the Cretaceous Frontier sandstone and Aspen formation. Table 4 presents data obtained from pebble counts in two conglomerate deposits in the Hoback Basin. The second conglomerate horizon studied, "undoubted Hoback formation," is the key horizon from which stratigraphic measurements were made (*see* Table 2 and Fig. 4). Although the exact source of many of the pebbles in that column was uncertain, the parent formations for those with determinable origins include the common Paleozoic and Mesozoic rocks still found in the surrounding region. Also, no recognizable Precambrian or Cenozoic (especially no "Pinyon-type") materials occur in this conglomerate. Size and shape data also indicate transportation was over short distances. This conglomerate within the Hoback formation can be distinguished from other conglomerate deposits in the area. The Pinyon and Pass Peak conglomerates are characterized by an abundance of well-rounded, pressure-marked Precambrian metaquartzites that are probably of distant Beltian origin, and by the low percentage or absence of Jurassic Twin Creek limestone fragments. The Camp Davis formation commonly includes Precambrian materials of local (Gros Ventre Range) origin. The absence from the Hoback formation of Precambrian materials of local derivation is related to the late (early Pliocene) rise of the Gros Ventre Range wherein the Precambrian is now exposed.

TABLE 4.—PEBBLE COUNTS IN HOBACK FORMATION CONGLOMERATES

	Deposit in Question*		Undoubted Hoback Formation†	
	Per cent of each size deformed	Per cent of total count	Per cent of total count	Per cent of each size deformed
Source Rock				
Cenozoic or Cretaceous				
"Salt and pepper" sandstone		8.1	0.0	
"Pinyon"-type pressure-marked metaquartzites such as found in the Pass Peak formation and Pinyon conglomerate		0.0	0.0	
Mesozoic				
Sandstone				
Stump formation—Late Jurassic		**	1.2	
Nugget formation—Early Jurassic		0.0	1.2	
Shale and siltstone, Triassic, formation uncertain		6.7	0.0	
Limestone, Twin Creek formation—Mid-Jurassic (equals "Lower Sundance")		5.4	19.7	
Paleozoic				
Sandstone and/or chert				
Phosphoria formation, Permian		1.4	0.0	
Tensleep sandstone, Pennsylvanian		14.8	5.8	
? Flathead quartzite, Cambrian		1.4?	0.0	
Precambrian		0.0	0.0	
Unidentified				
Quartzite or siliceous quartzitic sandstone, white to light gray		4.1	2.3	
Sandstone and/or chert		32.4	29.1	
Shale and siltstone		6.8	7.0	
Limestone and dolomite		18.9	33.7	
All sandstone and/or chert		58.1	37.3	
All shale and siltstone		13.5	7.0	
All limestone and dolomite		24.3	53.4	
All quartzite or siliceous quartzitic sandstone, white to light gray		4.1	2.3	
Shape				
Angular		24.3	0.0	
Subangular		42.8	41.9	
Subround		32.9	52.3	
Round		0.0	5.8	
Size (Maximum diameter in inches)				
½–1	33.3	12.2	37.9	0.0
1–2	25.0	37.8	39.1	2.9?
2–3	16.7	24.3	13.8	0.0
3–4	33.3	4.1	4.6	0.0
4–5	50.0	8.1	2.3	0.0
5–6	**	5.4	2.3	0.0
6–7	100.0	2.7	.....	.....

TABLE 4.—*Concluded*

	Deposit in Question*		Undoubted Hoback Formation†	
	Per cent of each size deformed	Per cent of total count	Per cent of total count	Per cent of each size deformed
Size (Maximum diameter in inches)				
7–8	**	2.7	.....	.....
8–9	**	0.0	.....	.....
9–10	**	0.0	.....	.....
Over 10	**	2.7	.....	.....
Maximum size in counting area	13 inches		6 inches	
Maximum size outside counting area	27 inches		16 inches	
	Cementation weak Deformation of particles associated with faulting, and folding No Precambrian or "Pinyon-type" materials		Cementation strong No faulting or folding. Particle deformation nearly zero No Precambrian or "Pinyon-type" materials	

* Concluded, on basis of this study, to be in the Hoback formation. Hillside west of Granite Creek-Little Granite Creek junction. Terminated by fault on south. (See Pl. 1, area B-4.) Counted area 2 by 3 feet. Total count 74.

† Conglomerate at a Paleocene level along U. S. Highway 187 opposite V-V Ranch. Area counted 2 by 3 feet. Total count 86.

** Indicates an item found elsewhere in the formation but not found in the counting area.

Data under the heading "Deposit in Question" were obtained from a pebble count in the conglomerate caught in a downfolded portion of the Cliff Creek thrust and affected by another fault that trends at an angle to that thrust. In the section on structural geology, this conglomerate is identified as part of the Hoback formation. The basis for this identification may be seen by comparing the data in the two columns of Table 4. In particular both conglomerates lack Precambrian and "Pinyon-type" material, and both are high in locally derived rock fragments. An interesting aspect of this second deposit is the high percentage of deformed pebbles and cobbles (Pl. 2, figs. 8–11). All particle sizes were affected. These deformed cobbles and pebbles are concentrated along the fault, and there seems to be no doubt that their condition resulted from crustal deformation. The materials in the very similar conglomerate shown in the right-hand column of Table 4 underwent almost no deformation, because they were not involved in faulting or tight folding. Similarly deformed cobbles were reported by Campbell (1906), but in the situation he described, the fracturing of boulders was not clearly due to crustal disturbance. However, his conclusion

> "... that such fracture can be produced only where the matrix is soft and flows under pressure, leaving the bowlders to take the brunt of the strain ..."

seems applicable, inasmuch as the matrix of the deposit in question certainly is soft.

SUMMARY AND CONCLUSIONS

This paper provides additional information concerning the vertebrate paleontology, stratigraphy, and structural relationships of the early Cenozoic Hoback formation in central western Wyoming. The sequence and times of occurrence of Laramide events are discussed in greater detail and with added certainty in the light of this new information.

Fossil mammals and other vertebrates have been found at several levels in the Hoback

formation. The newly discovered Battle Mountain local fauna includes a new species of phenacodontid condylarth, *Tetraclaenodon transitus*. The faunule occurs near the base of the formation and is of late Torrejonian (late middle Paleocene) age. Additions to the Dell Creek local fauna include a new species of mixodectid insectivore, *Elpidophorus minutulus*. Additional fossil mammals are reported from UM-Sub-Wy localities 7, 10, and 20.

Stratigraphic thickness, gross lithologic aspect, and major faunal levels are plotted against geologic time, revealing rates of sedimentary deposition for the Hoback formation and also showing the relationship between deposition rates and orogeny. A number of conclusions stem from this approach and from the earlier work of this author and other writers.

The Hoback formation is a synorogenic deposit composed of sedimentary material of local origin. Sediment was derived from mid-Laramide orogenic highlands to the west and accumulated in a rapidly subsiding trough which partially occupied the present site of the Hoback Basin. The formation exceeds 15,000 feet in thickness. About 12,620 feet of sediment were deposited in the geologically brief span of time between late middle Paleocene and earliest Eocene. Hence most of the formation accumulated at a rapid rate, although an initial acceleration and final deceleration in deposition rate are indicated by the dominantly fine-grained nature of the lowermost and uppermost portions of the formation. The time of beginning of deposition, changes in rate of deposition, and variations in gross lithologic aspect of the Hoback formation resulted from changing rates of erosion in mid-Laramide highlands. Erosion rates were in turn related to the times of beginning and culmination of the Middle Phase of the Laramide orogeny. This orogenic phase probably began in early Torrejonian (early middle Paleocene) time, and culminated sometime between the middle and end of the Torrejonian. Deposition of the Hoback formation was most rapid at or shortly after the culmination of orogeny, but continued at a decreasing rate until interrupted by thrusting of the First Stage of the Late Phase of Laramide orogeny in the late Wasatchian (late early Eocene).

During the time of deposition of the Hoback formation, the area was much lower, at an altitude of about 1000 feet. Deposition occurred in fluviatile, palludal, and lacustrine environments. The region was temperate, humid, probably heavily forested, and inhabited by a forest-dwelling mammalian fauna during the major part of the period of deposition. A savannah-type environment developed toward the close of the Paleocene or in the early Eocene.

The marginal Cliff Creek overthrust sheet is internally complex, with folds and faults parallel with and perpendicular to the thrust trace and in two sets at acute angles to the thrust trace. Folds in the overridden Hoback formation roughly parallel the thrust trace, locally pass beneath the overthrust plate, and are overturned eastward in the direction of thrusting. Early Cenozoic faulting and folding locally deformed cobbles and pebbles in the Hoback formation.

Conclusions reached in this paper, together with those of Eardley (1951, p. 287, 323, 325, and Fig. 215) summarized earlier in this paper, suggest that the sequence and timing of Laramide events in central western Wyoming and eastern Idaho was as follows:

Early Phase?—latest Cretaceous or early Paleocene. No evidence for this phase is recognized in the Hoback Basin.

Middle Phase—largely Torrejonian, culminating between the middle and end of the Torrejonian.

Late Phase

First Stage—late Wasatchian.

Second Stage—late Bridgerian or post-Bridgerian.

The phases and stages of orogeny were relatively brief and intense, although it cannot be shown that interorogenic times were completely anorogenic.

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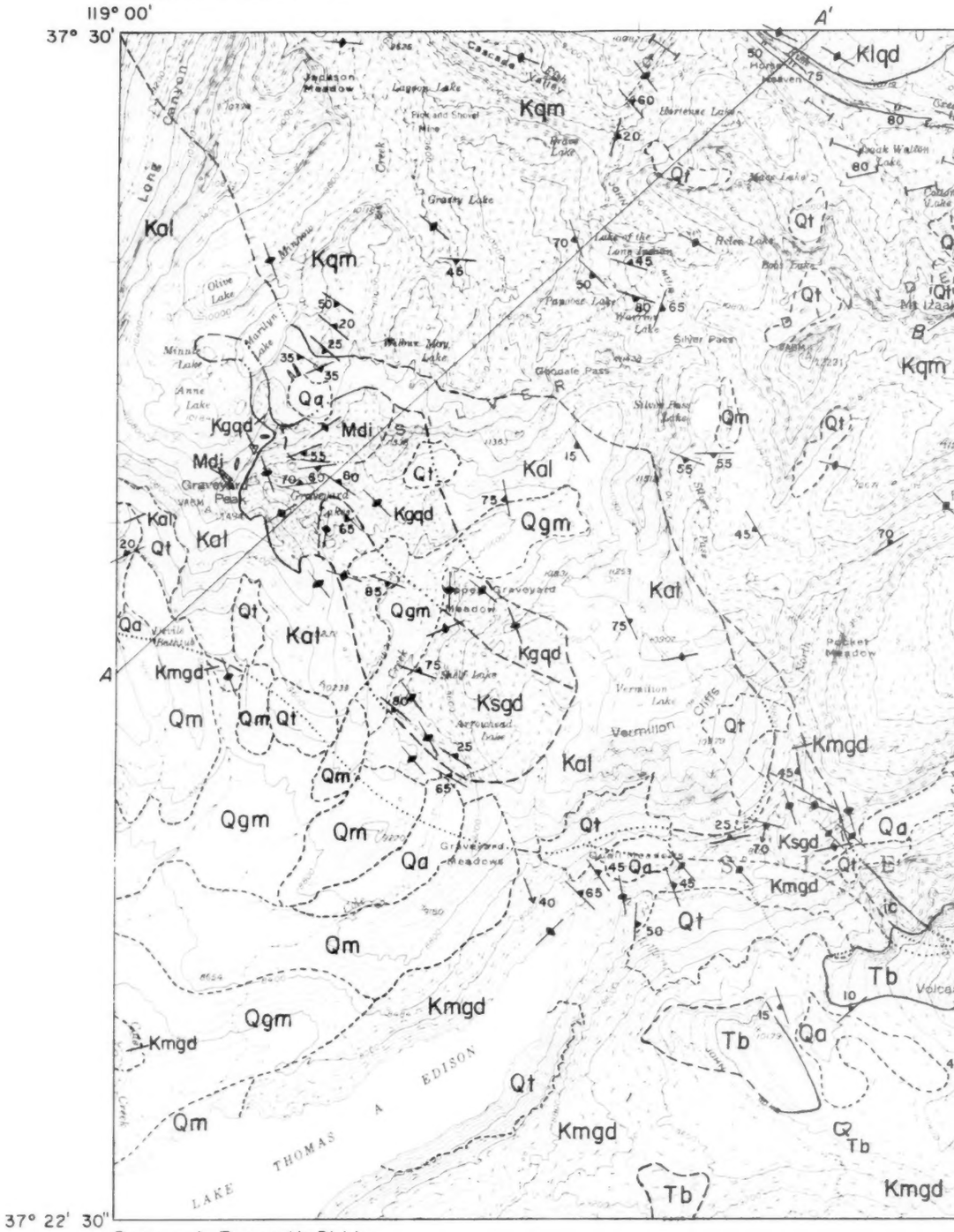

Base map by Topographic Division
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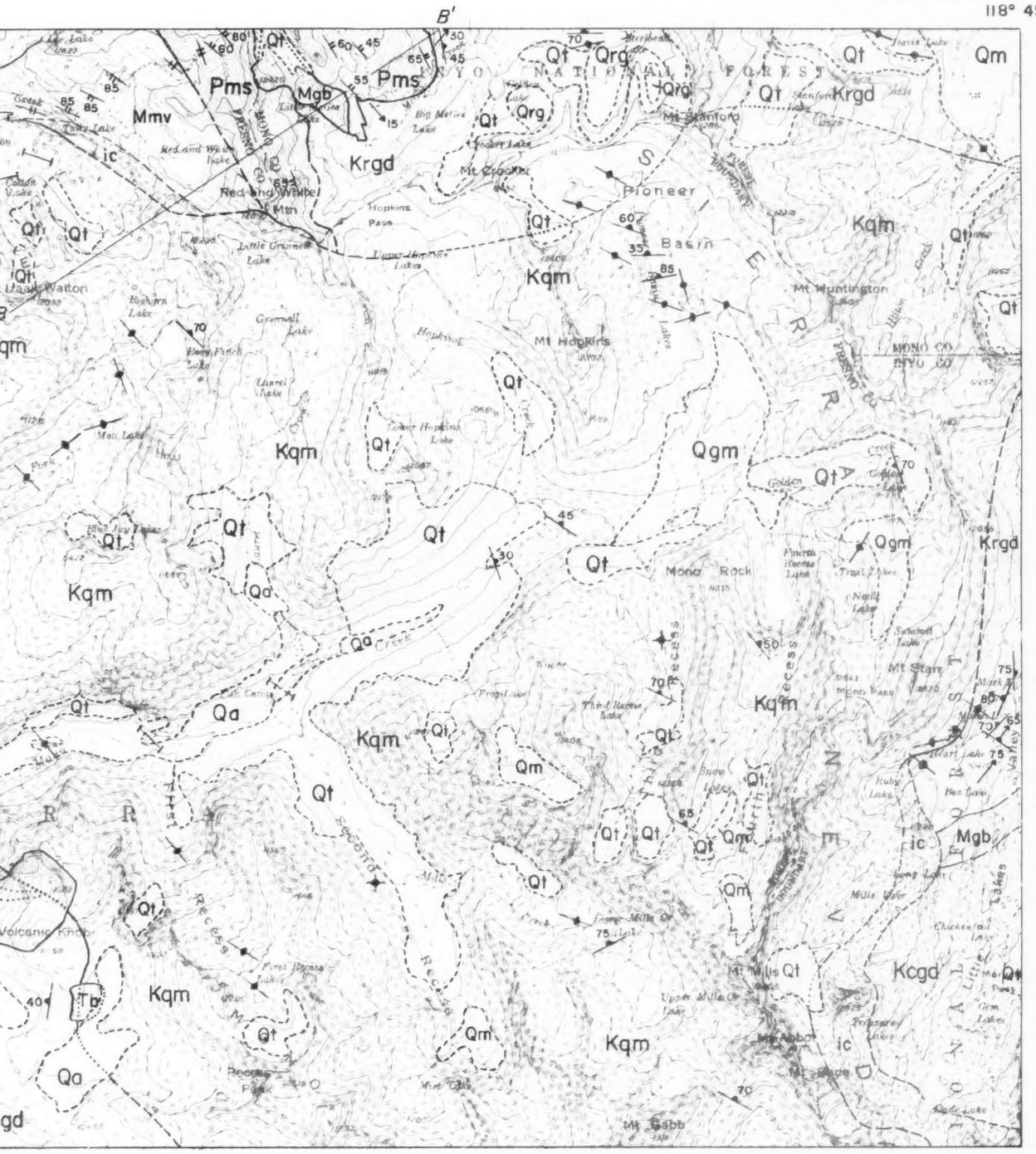
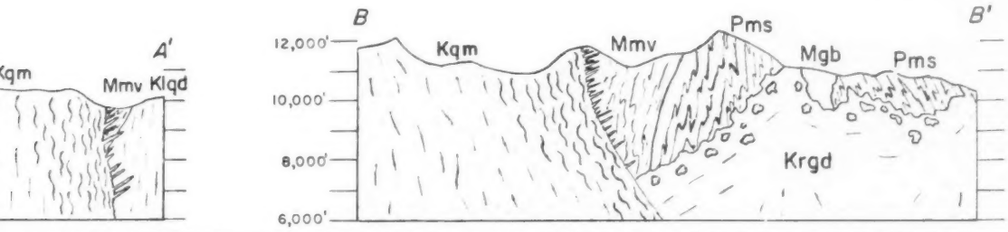


Geology by D. G. Sherlock and W. B. Hamilton, 1953–56



SECTIONS, NORTH HALF, MT. ABBOT QUADRANGLE, CALIFORNIA

Scale 1:62,500

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Contour interval 80 feet

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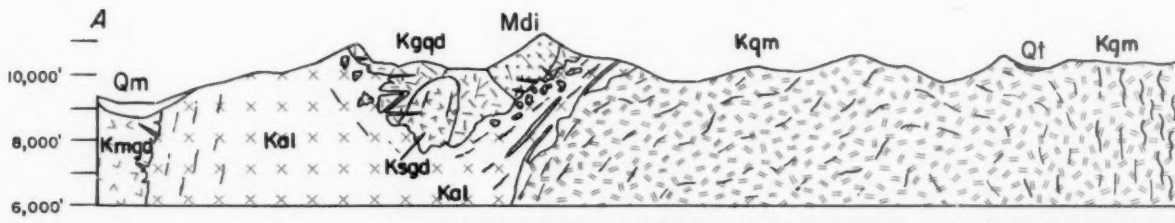

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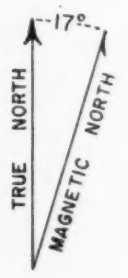

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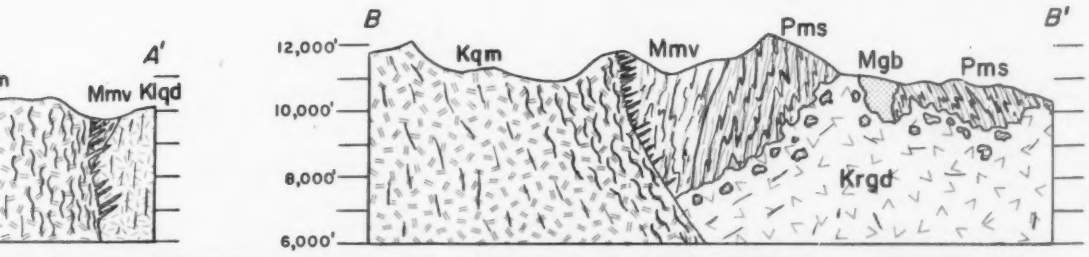

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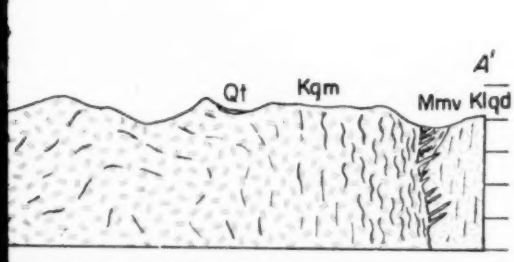
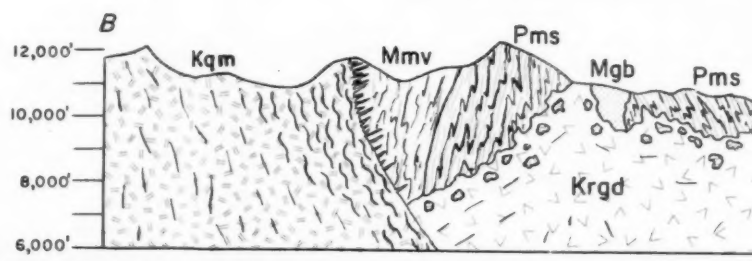




LOGIC MAP AND SECTIONS, NORTH HALF, MT. ABBOT QUADRANGLE, CALIFORNIA

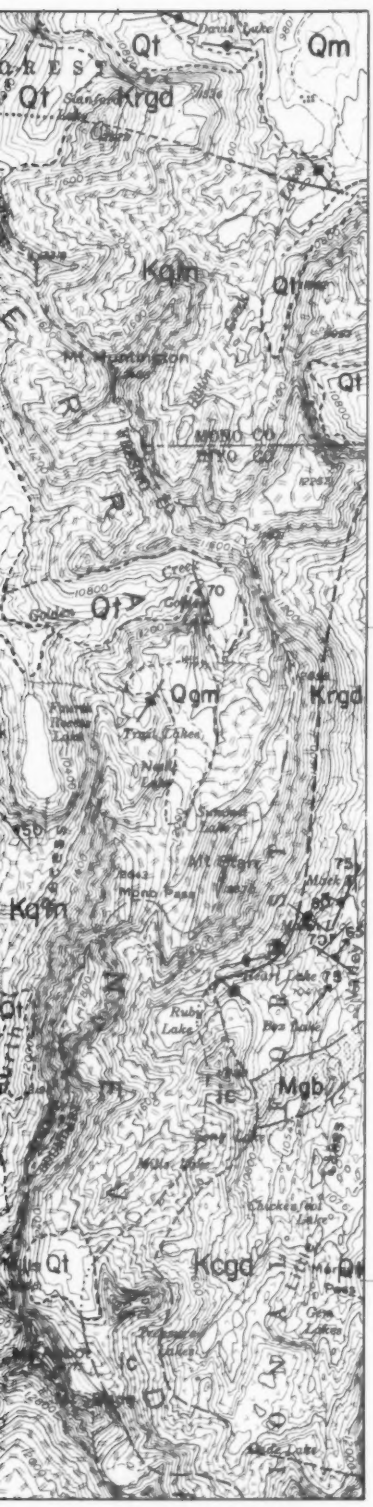
Scale 1:62,500

0 1 2 3 4 5 Miles

Contour interval 80 feet

Sherlock and Hamilton, Pl. 1

118° 45'



and W. B. Hamilton, 1953–56

NIA

QUATERNARY

- Qt — Talus
- Qa — Alluvium, outwash and ground moraine
- Qrg — Rock glacier
- Qm — Moraine
- Qgm — Ground moraine

TERTIARY(?)

- Tb — Basalt

CRETACEOUS

- Kqm | ic — Quartz monzonite of Mono Recesses; ic, injection complex
- Kal — Alaskite of Graveyard Peak
- Krgd — Granodiorite of Rock Creek
- Kcgd — Granodiorite of Chickenfoot Lake
- Klqd — Quartz diorite of Lee Lake
- Ksgd — Granodiorite of Shelf Lake
- Kmgd — Mt. Givens granodiorite
- Kgqd — Quartz diorite of Graveyard Lakes

MESOZOIC

- Mdi — Metadiorite
- Mgb — Metagabbro
- Mmv — Metavolcanic rocks

PALEOZOIC

- Pms — Metasedimentary rocks

- 80 — Protoclastic foliation *vertical, inclined*
- 45 — Primary planar flow structure *vertical, inclined, horizontal*
- 65 — Primary flow lineation
- 60 — Foliation of metamorphic rocks *vertical, inclined*
- Contact, accurately located
- Contact, approximately located
- Contact, inferred or surficial
- Contact, concealed

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GEOLOGY OF THE NORTH HALF OF THE MT. ABBOT QUADRANGLE, SIERRA NEVADA, CALIFORNIA

BY DONALD G. SHERLOCK[1] AND WARREN HAMILTON

ABSTRACT

The north half of the 15-minute Mt. Abbot quadrangle lies across the crest of the Sierra Nevada. The Cretaceous granitic rocks that underlie most of the area form eight large discordant plutons that range from quartz diorite to alaskite; the largest pluton is coarse prophyritic quartz monzonite. Pre-batholithic metasedimentary and metavolcanic rocks and metagabbro and metadiorite are minor.

The porphyritic quartz monzonite sent gently dipping dikes as much as 200 feet thick into its walls. These dikes in places make up more than half the height of exposed contact zones. A broad border zone of protoclastic flaser gneiss was formed in part of the pluton and in the adjacent wall rocks and dikes.

A contact between granodiorite and calcareous metasedimentary rocks swarms with dark inclusions. Their origin, puzzling in most places, is here clearly due to progressive hybridization of calc-silicate xenoliths; the changes take place within a few tens of yards of the contact. The xenoliths were amphibolitized concentrically and reconstituted to typical dark-inclusion texture and mineralogy and were drawn out from blocky xenoliths into spindle shapes.

Elsewhere, large irregular xenoliths of metadiorite were assimilated by alaskite. Within the resulting complex is a thick series of alaskite flow layers crowded with xenocrysts showing graded "bedding".

All the granitic plutons were intrusive; assimilation of mafic wall rocks by felsic magmas may have caused much of the diversity.

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INTRODUCTION

This paper describes the petrology and contact relationships of the granitic rocks in part of the Sierra Nevada batholith. Metamorphic rocks and Tertiary volcanic rocks are not extensive, and they and the surficial geology are treated but lightly. The Mt. Abbot quadrangle lies across the crest of the high central Sierra Nevada, a wilderness of ragged peaks, barren upland basins, and deep glaciated canyons. The north half of the quadrangle, 118 square miles, lies between the co-ordinates of 118°45′ and 119°00′ west longitude and between 37°22′30″ and 37°30′ north latitude. The geologic map (Pl. 1) includes an index map. The area is entirely within Sierra and Inyo

[1] Deceased.

National Forests and is mostly within the High Sierra Wilderness Area.

The Sierra crest, the drainage divide between the Owens Valley on the east and the San Joaquin Valley on the west, curves through the eastern part of the area (Pl. 2, fig. 1). Mt. Abbot attains the greatest altitude along the divide (13,715 feet); Mono Pass, crossed by the only trail over the crest within the area, is 12,000 feet high, as is McGee Pass, a few hundred yards north of the center of the area. From the crest, McGee, Hilton, and Rock (Little Lakes Valley) creeks flow northeast. Within the area their minimum altitude is about 10,000 feet.

West of the main Sierra crest, the topography is dominated by the valley of Mono Creek, trending west-southwest toward the South Fork of the San Joaquin River. The valley walls rise precipitously above the stream. East of the Second Recess, the U-shaped inner valley is 3000 feet deep and only 1½ miles wide (Pl. 3, fig. 3). The lowest altitude in the area, 7500 feet, is in the southwest corner. The Silver Divide (Pl. 3, fig. 1), with peaks more than 12,000 feet high and a low point at Silver Pass (10,600 feet), is about 3 miles north of and parallel to Mono Creek. Streams flowing northwest from the Silver Divide toward the Middle Fork of the San Joaquin are, at minimum altitude, at 8000 feet. Total relief in the area is greater than 6000 feet.

Topography was profoundly affected by repeated Quaternary glaciation, and except for the summits and a few ridges, the entire area was submerged by ice. Peaks are jagged, valleys steep-walled, and meadows and small lakes are abundant on the broad upland basins. Glacial debris was deposited in large quantities in the southwestern corner of the area, but rock surfaces, with or without a screen of rock rubble, dominate the remainder of the area. Many of the glacially carved cliffs have impressive talus slopes.

Like the rest of the High Sierra, the Mt. Abbot quadrangle receives heavy winter snows but little rain in summer; this climate makes field work pleasant and rewarding because of the insignificance of chemical weathering of the rocks. Four tiny fast-dwindling glaciers lie on the north slopes of 13,500-foot-plus peaks in the southeast corner of the area. Temperatures are cool even in summer, and frost occurs nightly except in the lower altitudes.

Timber line is near 11,000 feet, although trees are scarce over broad areas above 9500 feet because of the freshness of the glaciated rock surfaces.

There are no roads in the area, although the east margin is but a short distance from the end of the Rock Creek road, and the southwest corner is near the end of the Edison Lake road. The Mono Creek trail crosses the area from west to east, and the John Muir trail from north to south. The Fish Creek-McGee Creek trail, in and out of the north margin of the area, is another through trail. Most of the other trails are dead-end laterals; some of these are incorrectly located on the topographic map. It is easy to get around in the broad upper basins, but crossing from one basin to another can be dangerous, as the upper north slopes of even those saddles that look gentle on the topographic map may be at more than 45-degree angles.

The Southern California Edison Company completed in 1954 an earth dam across Mono Creek half a mile south of the mapped area, and the resulting Edison Lake extends 3 miles into the area. The dam is more than 4000 feet long, has a maximum height of 165 feet, and was constructed from the outwash gravels and moraines that floored the broad valley.

A complex vein of pegmatite, gray quartz, and red-stained quartz in the quartz monzonite near the northwest corner of the area contains a little magnetite and tiny quantities of pyrite, chalcopyrite, galena, and sphalerite. In 1953,

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PLATE 2.—CONTACT RELATIONSHIPS

FIGURE 1.—Sierra crest, looking S.30°W. across Long Lake. The injection complex of dike-layered rocks forms the left part of the left-hand peak, the triangular cliff below the skyline just left of center (*see* figure 1 for detailed sketch), and the right-hand peaks near the skyline; beyond is the quartz monzonite of Mono Recesses; granodiorite of Chickenfoot Lake is on the lower slopes.

FIGURE 2.—Sharp contact between granodiorite of Rock Creek and coarsely porphyritic quartz monzonite of Mono Recesses, by Ruby Lake; the quartz monzonite is the younger and near by truncates flow structures of the granodiorite. Knife is 3.5 inches long.

FIGURE 3.—Contact between granodiorite of Rock Creek and dark hybrid rock containing spheres of green epidote. Above Big McGee Lake.

FIGURE 1

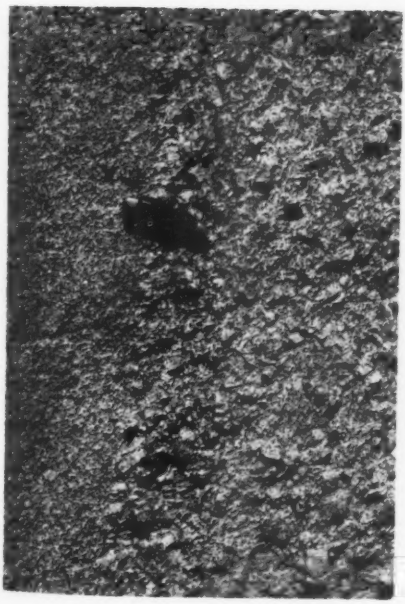

FIGURE 2

FIGURE 3

CONTACT RELATIONSHIPS

FIGURE 2

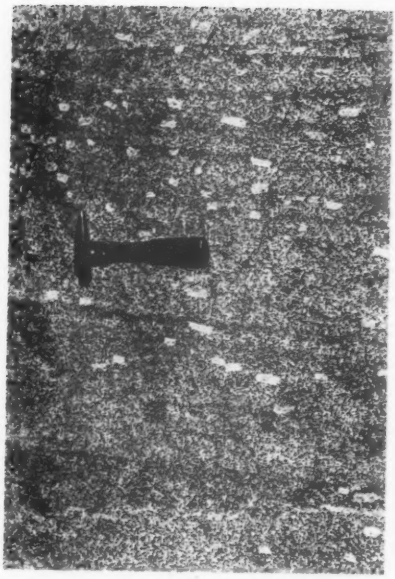

FIGURE 4

FIGURE 1

FIGURE 3

GRANITIC AND METAMORPHIC ROCKS

the vein was excavated as the Pick and Shovel Mine by W. A. Baldwin and G. T. Burns; a 30-foot inclined shaft was dug, but no ore was found. No other economic-mineral venture in the area is known, and no deposits of potential value were found. The most likely area for prospecting is that of metasedimentary rocks in the north part of the area, which might contain contact-metamorphic tungsten deposits.

Most of the field work was done by Sherlock during the summer of 1953 as the preliminary work for a thesis at the University of California. In June of 1954, prior to office work on the report, Sherlock was killed in the course of other field work while attempting to ford a meltwater-swollen river in western Idaho. The project had been entirely Sherlock's, but Hamilton decided to complete it and visited the area briefly in 1954. Largely from the map, field notes, hand specimens, and thin sections left by Sherlock, Hamilton prepared a preliminary report. The need for the further field work which Sherlock had intended to do was apparent. In 1956 Hamilton revisited the area, studying particularly contact relationships. About 120 thin sections and 300 hand specimens, two-thirds of them collected by Sherlock, were studied. Photographs are by Hamilton.

Contacts shown by solid lines on the geologic map (Pl. 1) are in general accurate within several hundred feet. Because Sherlock did not distinguish between accurate and approximate contacts on his field map, all contacts mapped by him are shown as dashed lines, although many must warrant solid lines. Some dashed bedrock contacts may be incorrectly located by as much as half a mile in remote places.

Most of the surrounding area has been mapped geologically by others. Rinehart and Ross (1956; 1957) have mapped the Mt. Morrison quadrangle to the north and the Casa Diablo Mountain quadrangle to the northeast. Rinehart has mapped the Devils Postpile quadrangle to the northwest, and Bateman (1956) the Mt. Tom quadrangle to the east. The south half of the Mt. Abbot quadrangle has been mapped by P. A. Lydon. The Huntington Lake area mapped by Hamilton (1956a; 1956b) almost meets the area from the southwest. Reconnaissance by Mayo (1941) of the High Sierra region included this area. Birman (1954) and Webb (1956) have studied the surficial geology of parts of the area.

Information given generously by P. C. Bateman, D. C. Ross, C. D. Rinehart, and P. A. Lydon about the margins of the area, and comments by them and by E. B. Mayo on the manuscript, have been invaluable and are acknowledged gratefully. Sherlock worked alone in the area; Hamilton was accompanied by G. J. Neuerburg in 1954 and by G. J. Klapper in 1956.

The plutonic rocks are named on the basis of their mineral compositions. Gabbro has plagioclase more calcic than An_{50} and diorite has plagioclase more sodic than An_{50}, and neither contains more than 10 per cent quartz. The granitic rocks contain more than 10 per cent quartz and are named for their feldspar ratios, the arbitrary boundaries between quartz diorite, granodiorite, quartz monzonite, and granite being placed at the ratios of plagioclase to potassic feldspar of, respectively, 9:1, 5:3, and 3:5. Alaskite is leucocratic granite with less than 2 per cent (and generally less than 1 per cent) of dark minerals.

A pluton is a separate intrusive mass, generally bounded sharply against its neighbors, of any size from an irregular mass covering

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PLATE 3.—GRANITIC AND METAMORPHIC ROCKS

FIGURE 1.—Silver Divide, looking S.55°W. from Sierra crest at north edge of area. Foreground ridges are metatuff, and background is quartz monzonite of Mono Recesses.

FIGURE 2.—Metasedimentary rocks north of Little McGee Lake, from the northwest. The contortion of the dark schist and light siliceous hornfels is shown by the discontinuous layering. A dark dike cuts the light rocks above the center of the picture.

FIGURE 3.—Mono Creek Valley, looking N.25°E. across mouth of First Recess. The glacial trough is cut parallel to northeast joints in quartz monzonite of Mono Recesses; talus mantles the lower slopes. The dark tip of Red and White Mountain, left of center, is of metamorphic rocks.

FIGURE 4.—Dark streaky quartz monzonite with oriented phenocrysts. The dark minerals show no flowage around the rectangular potassic feldspars, and the phenocrysts are poikilitic, so the phenocrysts are probably of replacement origin. Hammer is 10 inches long; complex of Graveyard Peak, 1¼ mile northeast of Graveyard Peak, 0.6 mile N.18°E. of peak 11,227.

less than 1 square mile to a continuous body cropping out over several hundred square miles. Names have been given to the various granitic plutons for convenience in describing them, but these names are not of formal stratigraphic nomenclature. The plutons are named for geographic features within the area except for the granodiorite of Rock Creek, named by Rinehart and Ross (1957), and the Mt. Givens granodiorite, named Mt. Givens pluton by Hamilton (1956a).

Outline of Geology

The northern half of the Mt. Abbot quadrangle is within the Sierra Nevada portion of the vast California batholith of late Mesozoic age. The western zone of this central Sierra Nevada section of the batholith is mostly quartz diorite, although much of the rock is more mafic and intrudes low-grade metamorphic rocks dominated by mafic metavolcanic rocks. Granodiorite is the most abundant type in the medial zone, where metamorphic rocks are sparse. The eastern zone is dominated by quartz monzonite, and the large areas of metamorphic rock are dominantly metasedimentary rock, with lesser metavolcanic rocks, chiefly felsic. Although the batholith is composed of hundreds of separate intrusive plutons of different types of granitic rocks, these longitudinal zones of different dominant types are obvious. Both batholith and wall rocks thus have broadly parallel changes in composition across the batholith. Hamilton (in press) has described the geology of the batholith and compiled a geologic map across the batholith through the Mt. Abbot quadrangle.

The north half of the Mt. Abbot quadrangle is mostly within the eastern quartz monzonitic zone of the batholith, but the southwest corner extends into the medial granodioritic zone. Metamorphic rocks are restricted to the north edge of the area, which contains the south end of a large mass of Paleozoic metasedimentary rocks and Mesozoic(?) metavolcanic rocks. Next oldest are small metagabbro and metadiorite masses. Most of the area is formed of granitic rocks, dominated by the youngest(?) pluton, a great northwest-trending body of porphyritic quartz monzonite with granite facies. The seven other plutons mapped include three which are mostly of quartz diorite, three of granodiorite, and one of alaskite. Stoping, assimilation, forceful injection, and perhaps zone melting were important mechanisms of emplacement of the granitic plutons. There is no evidence for large-scale metasomatic granitization.

Among the features of interest in the area are the protoclastic border zone of porphyritic quartz monzonite against metavolcanic rocks, dark inclusions clearly formed by hybridization of metasedimentary rocks in granodiorite, and a remarkable assimilation complex of alaskite, quartz monzonite, and metadiorite.

Several small areas of Tertiary(?) basalts lie on granitic rocks in the south part of the area. Quaternary sediments are dominated by talus and glacial debris.

Metamorphic Rocks

The north-central part of the area is underlain by the south end of a large pendant of metamorphic rocks. The eastern half of this pendant is of metasedimentary rocks that are at least in part Paleozoic. The western half is of metavolcanic rocks that are probably Mesozoic. The rocks are isoclinally deformed with northwest strikes and steep dips. Superimposed on the effects of low-grade regional metamorphism are those of moderately intense contact metamorphism. Small masses of metamorphic rocks, not differentiated on Plate 1, lie within the injection complex in the southeast corner of the area.

METASEDIMENTARY ROCKS OF MAIN PENDANT: Schist and siliceous hornfels are dominant, and the dark schists are in marked contrast to the white siliceous hornfelses (Pl. 3, fig. 2). Marble is present only in the east part of the pendant near the north edge of the area, but calc-silicate rocks are widespread.

The schists are medium to dark gray where fresh and reddish brown where weathered; in most cases their grain size is less than 0.1 mm. The rocks commonly have 15–25 per cent orange-brown biotite, more quartz than plagioclase, and small amounts of muscovite, epidote, and actinolite or hornblende. The rocks become slightly coarser toward the contacts with granitic rocks, but schistosity is retained.

The siliceous hornfelses are dense and almost aphanitic, white or light greenish gray; quartz is the dominant mineral. Different combinations and amounts of epidote, clinozoisite, actinolite, diopside, muscovite, and plagioclase are present. Some of these rocks are almost entirely quartz. Grain size is commonly about 0.1 mm, and the quartz is moderately flattened.

The marbles are pure, coarse-grained, and gray, some with markedly flattened calcite

grains typically 2–3 mm long, some with equant anhedra that are more than 1 cm across. The calc-silicate hornfelses contain plagioclase and diopside and generally subordinate actinolite, hornblende, cordierite, and biotite.

Rinehart and Ross (1954) studied in detail the pendant to the north, where the intensity of deformation and of contact metamorphism was less. They found a lower 7000-foot-thick unit of argillite, locally containing Ordovician graptolites; a middle unit, 10,000 feet thick, of siliceous hornfels, calcareous quartzite, and marble, containing productid brachiopods of probable Pennsylvanian age near the top; and an upper unit of siliceous hornfels containing poorly preserved brachiopods. The metasedimentary rocks of the Mt. Abbot quadrangle are on strike with rocks stratigraphically above the fossiliferous Pennsylvanian(?) limestone (D. C. Ross, personal communication) and are presumed to be late Paleozoic.

METAVOLCANIC ROCKS OF MAIN PENDANT: The west part of the pendant of metamorphic rocks is of metavolcanic rocks in which tuffs are dominant over flows, and which range in composition from andesite to rhyolite. Most rocks are greenish. Most of the flow rocks are but weakly schistose, and their volcanic textures are generally well preserved in the metamorphic mineral assemblages. Relic phenocrysts of bipyramidal quartz spot the metarhyolites. Altered phenocrysts of potassic feldspar (now muscovite), of plagioclase (now albite), and of mafic minerals (now hornblende with or without biotite), are abundant. Volcanic-groundmass textures are preserved crudely by laths of albite, greatly recrystallized, and finely granular albite, quartz, muscovite, clinozoisite, hornblende, and biotite.

The metatuffs are slaty or phyllitic, their lapilli greatly sheared and flattened into thin lenses. Initial size of lapilli ranged from less than 1 mm to half a meter but fragments a few millimeters across were most common. The clasts show up clearly in hand specimen, as the tuffs are varicolored in light to dark greenish, purplish, and grayish hues. The biotite of the metavolcanic rocks is typically olive green in contrast to the orange-brown biotite of the metasedimentary rocks.

The effect of contact metamorphism on the rocks, except the coarsening toward the granitic rocks, is obscure. Great marginal shearing by the quartz monzonite of Mono Recesses has complicated the metamorphic effects.

No fossils are known from the metavolcanic rocks of the area. In the similar rocks of the Devils Postpile quadrangle to the northwest, C. D. Rinehart found several clams. Although specific determinations have not yet been made, the fossils indicate a Mesozoic age (D. C. Ross, personal communication). The metavolcanic rocks of the Mt. Abbot quadrangle are also considered to be Mesozoic.

OTHER METAMORPHIC ROCKS: Mafic metamorphic rocks are widespread among the wall rocks of the marginal injection complex of the quartz monzonite of Mono Recesses in the southeast corner of the area. These metamorphic rocks were not mapped, but they and the associated metagabbro form a discontinuous screen between the quartz monzonite and the granodiorites of Rock Creek and Chickenfoot Lake. The metamorphic rocks include biotite and chlorite-biotite schist, amphibolite, and calc-silicate rock. At least some of the amphibolite is of metasedimentary origin. Near the east end of Ruby Lake, large smoothly tapered xenolithic lenses of amphibolite (as much as 200 feet long and 20 feet thick) in the injection complex contain abundant small remnants of calc-silicate hornfels.

DIKE ROCKS: Throughout the area the metamorphic rocks are cut by dikes, typically a few feet thick, which commonly are sheared and partly reconstituted to new mineral assemblages. Some are greatly altered. The primary mafic minerals of a quartz diorite dike near the northern metagabbro were reconstituted to mottled granular hornblende and a little brown biotite, plagioclase was partly granulated and greatly altered, and quartz was thoroughly crushed. A biotite-hornblende granodiorite dike in metasedimentary rock near the head of Fish Creek was crushed to flaser gneiss but was only slightly reconstituted.

The metavolcanic rocks near the protoclastic flaser gneisses of the border of the quartz monzonite of Mono Recesses were injected by abundant thick dikes from the quartz monzonite. The dikes are aplite, pegmatite, and quartz monzonite, and many are crushed. The dikes form a crudely layered complex that dips at a moderate angle toward the quartz monzonite.

METAMORPHISM: Prior to the intrusion of the granitic rocks, the sedimentary and volcanic rocks were deformed isoclinally and reconstituted, presumably at the same time, to mineral assemblages of the greenschist facies. Granitic plutons and dikes cut the structures at all angles. Dikes crosscutting isoclinal structures were in turn variably sheared and reconstituted even in the interior of the meta-

TABLE 1.—MINERALOGIC COMPOSITIONS OF PLUTONIC ROCKS OF NORTH HALF OF MT. ABBOT QUADRANGLE
(Volumetric percentages generalized.)

Pluton	Quartz	Plagioclase		Potassic feldspar	Biotite	Hornblende	Augite
		amount	composition				
Quartz monzonite of Mono Recesses	20–35	25–40	An_{15}–An_{30}	25–50	1–5	0–2	
Alaskite of Graveyard Peak	20–40	15–30	An_{5}–An_{20}	40–65	tr.–2		
Granodiorite of Rock Creek	15–35	25–70	An_{25}–An_{40}	2–40	3–10	tr.–10	
Granodiorite of Chickenfoot Lake	15–30	35–65	An_{30}–An_{40}	10–30	3–12	1–7	
Quartz diorite of Lee Lake	10–20	65–80	An_{25}–An_{35}	1–5	6–12	0–4	
Granodiorite of Shelf Lake	7–20	50–70	An_{35}–An_{50}	7–15	5–13	10–15	0–1
Mt. Givens granodiorite	10–25	40–70	An_{30}–An_{40}	10–25	2–10	2–10	
Quartz diorite of Graveyard Lakes	8–15	45–70	An_{40}–An_{45}	3–15	5–17	6–15	
Metadiorite (near Graveyard Lakes)	3–10	55–75	An_{40}–An_{45}	0–2	8–20	7–20	tr.–5
Metagabbro, northern mass	1–5	45–55	An_{50}–An_{70}	1–3	2–5	35–45	
Metagabbro, eastern mass	1–2	50–65	An_{40}–An_{65}	1	5–15	20–40	0–10

morphic mass, so some metamorphism followed the main deformation. It is not possible to distinguish clearly between the effects of regional and contact metamorphism in the metamorphic rocks, and the two may not have been separated sharply in either time or process. The mineral assemblages due to contact metamorphism probably belong mostly to the cordierite-anthophyllite subfacies of the amphibolite facies of Turner (Turner, 1948; Turner and Verhoogen, 1951).

GABBROIC AND DIORITIC ROCKS

A small metagabbro mass lies in the southeast part of the area, and another in the pendant

PLATE 4.—PHOTOMICROGRAPHS

FIGURE 1.—Metagabbro, above Little McGee Lake. Aggregates of granular hornblende (patchy), with a little biotite, darken toward their margins and are probably secondary after pyroxene. Plagioclase photographs white. Ordinary light.

FIGURE 2.—Metadiorite, 0.5 mile northeast of Graveyard Peak. Augite (upper right) is rimmed by hornblende. Dark biotite is right, and magnetite left, of center. Ordinary light.

FIGURE 3.—Oscillatory-zoned andesine, granodiorite of Rock Creek, Pioneer Basin. Crossed nicols.

FIGURE 4.—Inclusions of plagioclase, with sodic reaction rims, in phenocryst of potassic feldspar. Near-by inclusions are optically continuous. Quartz monzonite of Mono Recesses, near Goodale Pass. Crossed nicols.

FIGURE 5.—Coarsely perthitic microcline. Alaskite of Graveyard Peak, Long Canyon; crossed nicols.

FIGURE 6.—Amphibolite rim of calc-silicate xenolith. Granular hornblende and clinopyroxene (left) were replaced by big poikilitic anhedra of hornblende (right). Plagioclase (white) is more coarsely granular on the right. Mt. Givens granodiorite, below high-water level at upper end of Lake Edison; ordinary light. Outcrop photograph given by Figure 4 of Plate 5.

PLATE 5.—XENOLITHS

FIGURE 1.—Xenoliths of metagabbro in granodiorite of Rock Creek at Box Lake. The variety of textures and compositions is greater than that of the near-by metagabbro.

FIGURE 2.—Contact between calc-silicate hornfels, and granodiorite of Rock Creek with xenoliths. The hornfels blocks have amphibolitic margins. North-northeast 0.7 mile from Big McGee Lake. See figures 5 and 6 of Plate 6.

FIGURE 3.—Hybridized xenoliths 50 feet into granodiorite from figure 2. Calc-silicate mineralogy has been completely modified.

FIGURE 4.—Edge of large calc-silicate xenolith in Mt. Givens granodiorite, with half-inch layer of dark amphibolite along contact, and small hybridized xenoliths in granodiorite. See figure 6 of Plate 4.

FIGURE 1

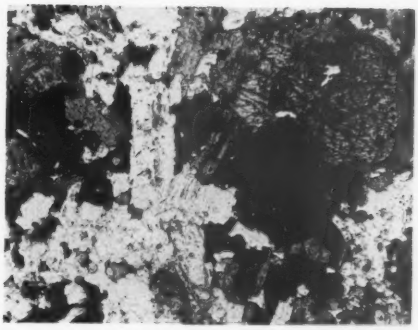


FIGURE 2

FIGURE 3

FIGURE 4

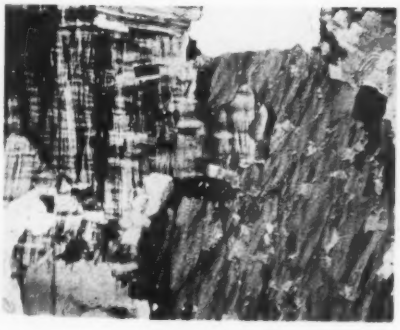


FIGURE 5

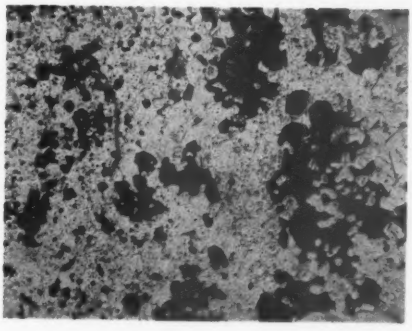


FIGURE 6

PHOTOMICROGRAPHS

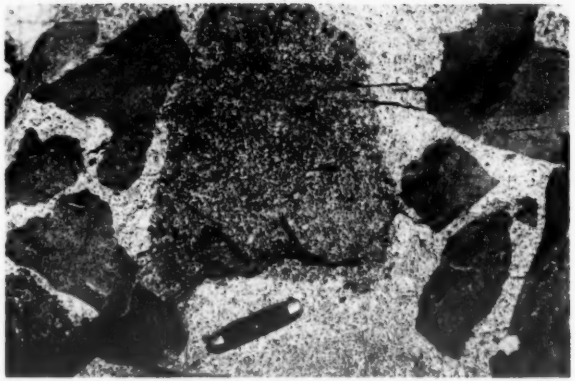


FIGURE 1

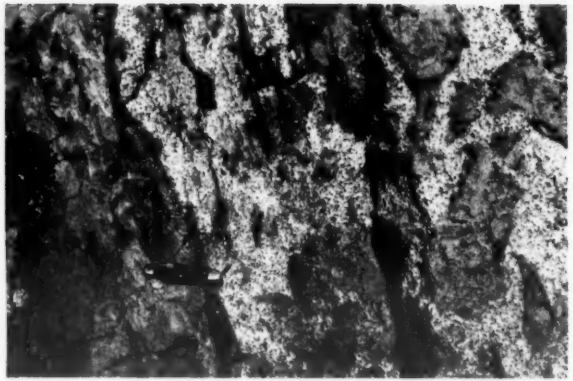


FIGURE 2

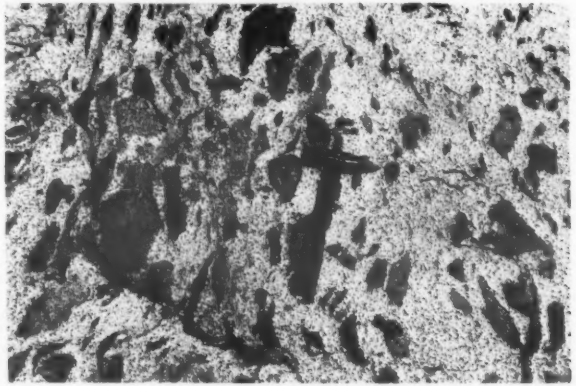


FIGURE 3

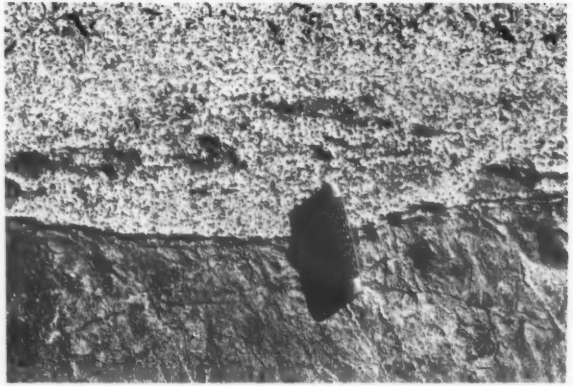


FIGURE 4

XENOLITHS

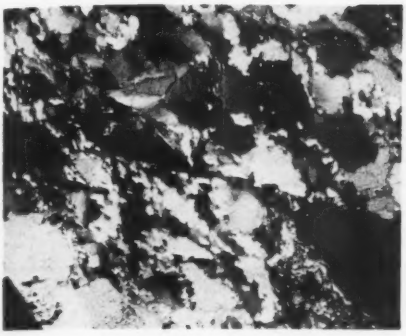


FIGURE 1

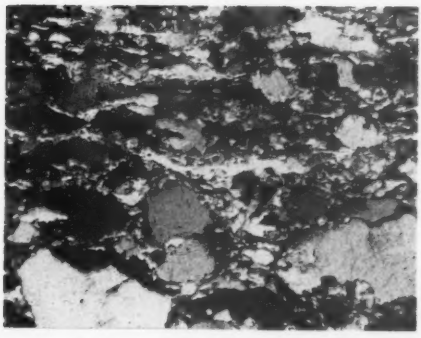


FIGURE 2

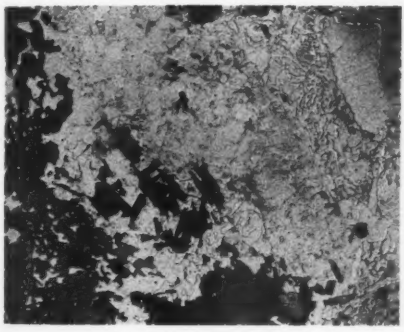


FIGURE 3

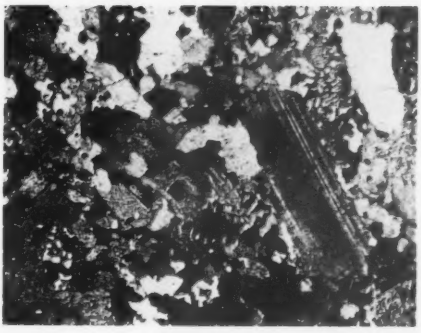


FIGURE 4

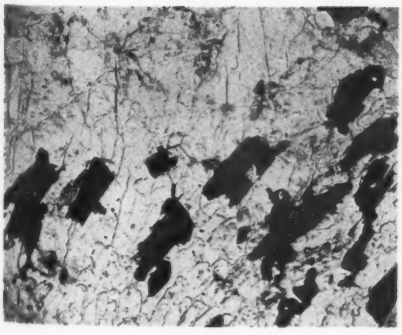
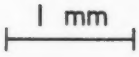


FIGURE 5

FIGURE 6



PHOTOMICROGRAPHS

FIGURE 1

FIGURE 2

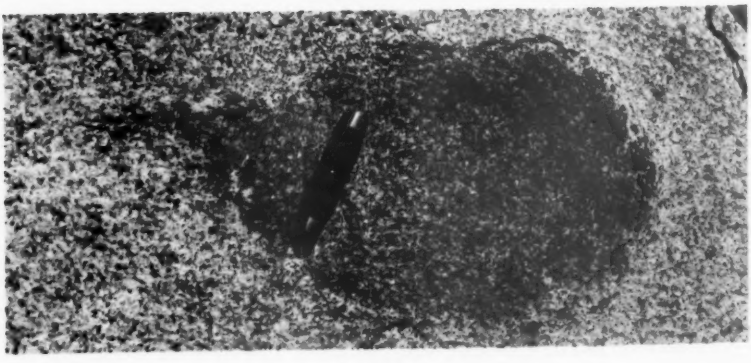


FIGURE 3

COMPLEX OF GRAVEYARD PEAK

of metamorphic rocks to the north. More metagabbro, associated with mafic metamorphic rocks, lies within the nonsubdivided injection complex between the quartz monzonite of Mono Recesses and the granodiorite of Chickenfoot Lake. A small metadiorite mass lies to the west. These rocks are so unlike the granitic rocks as to require separate treatment. Mineralogic compositions are summarized in Table 1.

NORTHERN METAGABBRO: Within the metasedimentary rocks near the north edge of the area is a lenticular mass 1 mile long and ½ mile wide of hornblende metagabbro, composed of somewhat more labradorite than hornblende. The rocks are medium-grained and medium light to medium dark gray; the color depends more on the plagioclase, which is light to dark, than on the amount of hornblende. Compositional changes are gradual and are inconspicuous in the field. Labradorite is in subhedra 1–2 mm long, is much altered, and contains abundant tiny inclusions of hornblende, magnetite, epidote, and muscovite; new growths of clear and more sodic plagioclase are present on some grains. The hornblende is greenish black in hand specimen and green in thin section and is mostly in 1–2 mm aggregates of anhedra smaller than 0.1 mm. Both separate grains and whole aggregates show crude color zoning, the color darkening and becoming slightly bluish outward. The rocks show slight crushing, with a little bending and marginal granulation of plagioclase, but no foliation. The over-all texture (Pl. 4, fig. 1) suggests an origin as a semidiabasic pyroxene gabbro in which the pyroxene has been completely altered to hornblende. The hornblende has in turn been partly altered to brown biotite, which is in tiny grains strewn through and about the hornblende, in aggregates of small anhedra, and in large anhedra. The rocks contain a few per cent of interstitial orthoclase and strained quartz, which are not intergrown. Sphene is plentiful as anhedra within hornblende and along cleavage folia of biotite and as rims on magnetite. ("Magnetite" in this paper includes both ilmenite and true magnetite, which were not distinguished from each other.)

The metagabbro is massive, and its outcrops are shaped like those of granitic rocks. Inclusions of still darker rocks are locally present in it, as are small patches of hornblendic migmatite.

The metagabbro was intruded by the granodiorite of Rock Creek, and its metamorphism—partial reconstitution, without pervasive shearing—was probably a contact effect of the granodiorite. Contacts between metagabbro and metasedimentary rocks were not seen, but as the latter are sheared pervasively and are cut by dark dikes, it is likely that the gabbro was emplaced after the regional metamorphism.

EASTERN METAGABBRO: A mass of hornblende metagabbro one-third of a mile wide extends 1 mile into the southern part of the area from

PLATE 6.—PHOTOMICROGRAPHS

FIGURE 1.—Flaser gneiss developed from porphyritic quartz monzonite of Mono Recesses, near Hortense Lake. This is in the protoclastic border zone against metavolcanic rocks. Crossed nicols.

FIGURE 2.—Crushed aplite, in dike from quartz monzonite of Mono Recesses which cuts metavolcanic rocks. Fish Creek, at creek from Lee Lake; crossed nicols.

FIGURES 3, 4.—Contaminated granodiorite of Chickenfoot Lake, at contact with metagabbro at Long Lake. Figure 3, ordinary light, figure 4, crossed nicols, same field. A poikilitic xenocryst of hornblende is in the lower left. The mosaic and intergrowth textures of quartz and feldspar are distinctive of this contact facies.

FIGURES 5, 6.—Contact between granodiorite of Rock Creek (upper left) and hybridized calc-silicate xenolith. Figure 5, plane light, figure 6, crossed nicols, same field. A large poikilitic potassic feldspar (light gray, figure 6) extends across the contact and has replaced much of the material of the schistose xenolith. One foot from granodiorite contact of figure 2 of Plate 5.

PLATE 7.—COMPLEX OF GRAVEYARD PEAK

FIGURE 1.—Graveyard Peak (right), from east-northeast. Rock types: Kal, alaskite; Kgqd, quartz diorite; Mdi, metadiorite.

FIGURE 2.—Layers of contaminated alaskite, with gravity-graded dark-mineral content, overlie and truncate the planar flow structure, which dips moderately right, of porphyritic quartz monzonite. Contaminated border facies of alaskite of Graveyard Peak, near main contact with quartz monzonite of Mono Recesses, northeast 0.6 mile from peak 11,227.

FIGURE 3.—Xenolith of metadiorite in contaminated alaskite. Mafic reaction-xenocrysts increase toward the xenolith; north-northeast 0.4 mile from peak 11,227.

its east margin. Like the northern metagabbro, this mass contains more plagioclase than hornblende; the eastern mass, however, has been much more thoroughly recrystallized.

The rock is medium- to coarse-grained and gray, the darkness of the gray varying with both color of plagioclase and amount of mafic minerals. Hand-specimen appearance is illustrated by the xenoliths of figure 1 of Plate 5. Plagioclase is partly in subhedra and partly in equant anhedra, mostly about 1 mm but some as large as 5 mm. Some plagioclase is rather uniform, generally near An_{50}; some is zoned—an extreme range seen in one section was An_{80} to An_{25}. Albite and acline twins are common. (As they are generally indistinguishable by ordinary petrographic study, pericline and acline twins are here considered under the latter name.) Hornblende is green and is characteristically in poikilitic anhedra 1–3 mm long, which are in many places discontinuous in the plane of thin section. Diopsidic augite rimmed by hornblende is present in some specimens. Biotite is in ragged olive-brown anhedra, mostly secondary after hornblende, with interlaminae of muscovite and chlorite. Quartz and orthoclase are interstitial. Epidote is common in veins and disseminated grains.

The mass is broadly uniform, although abrupt local variants differ greatly from the typical medium-grained rock. The most conspicuous variant is at the south contact of the mass at Long Lake, where 15 feet of highly variable metagabbro lies between normal metagabbro and granodiorite, sharply bounded against each. Grain size in this rock ranges irregularly from 1 to 5 mm, and composition varies greatly within hand specimens. The rock is green because much of the hornblende has been altered to chlorite and actinolite.

The metagabbro is cut by abundant thick gently dipping aplite and pegmatite dikes which increase in number westward toward the quartz monzonite of Mono Recesses; the metagabbro contacts continue into the marginal injection complex of that pluton but are not shown on the geologic map within that zone of extreme diking. In the metagabbro within a fraction of an inch of a dike, most of the hornblende was altered to biotite and chlorite, and quartz and orthoclase were added.

Xenoliths of metagabbro are abundant in the granodiorite of Rock Creek near the contact (Pl. 5, fig. 1) but are virtually lacking in the granodiorite of Chickenfoot Lake. Both granodiorites show the effect of marginal contamination by assimilation of metagabbro. The eastern metagabbro was thoroughly recrystallized, presumably by the contact action of the flanking granodiorites. Gabbroic texture is preserved chiefly in the shape and zoning of some of the plagioclase. The augite may be relic from the initial mineral assemblage; the poikilitic hornblende, however, shows extreme recrystallization. The eastern metagabbro, like the northern, is broadly uniform and lacks metamorphic foliation or compositional layering, and probably originated as an intrusive mass.

OTHER METAGABBRO: Within the dike-injection complex along the southeast margin of the quartz monzonite of Mono Recesses is much metagabbro similar in appearance to that of the eastern metagabbro. The masses of this other metagabbro were neither mapped nor studied petrographically. Other host rocks in this injection complex are mafic metamorphic rocks and the granodiorites of Rock Creek and Chickenfoot Lake.

METADIORITE: A small mass of mafic augite-biotite-hornblende metadiorite crops out in an area ½ by 1 mile in the north part of the plutonic complex of the west part of the area. Changes within the mass are slight and gradational. A gradationally bounded vertical dike of very similar rock, several tens of feet thick, is exposed within the mass in the saddle 0.6 mile southeast of Marilyn Lake.

The metadiorite is medium light gray with black mafic minerals and light-gray 1-mm plagioclase. The rock is composed of weakly zoned calcic andesine, biotite, and hornblende; the latter in many cases contains cores or small remnants of augite (Pl. 4, fig. 2). The outlines of the mafic minerals conform to those of the well-shaped plagioclase grains. Both biotite and hornblende are so poikilitic that they are in many cases discontinuous within a section; exceptionally, biotite sieves are 1 cm in diameter and hornblende half that, although separate areas of "solid" mineral in these crystals may be only 1 mm in diameter. Less widespread phases of the metadiorite contain compact anhedra of biotite and hornblende of about the same size as the accompanying subhedral plagioclase. Still other phases contain seriate plagioclase and mafic minerals. Augite is nearly colorless in thin section, hornblende is green, and biotite is olive brown. The minor quartz and orthoclase are in small interstitial grains. The textures indicate clearly that hornblende replaced augite and that biotite replaced hornblende. This mineralogic reconstitution and the poikiloblastic texture probably resulted from contact metamorphism by the adjacent younger plutons.

The metadiorite is cut by dikes, most of which are gently dipping, from the adjacent alaskite of Graveyard Peak. The north contact of the mafic mass is the gradational upper limit of the remarkable assimilation complex described on later pages. The contact between metadiorite and the quartz diorite of Graveyard Lakes was not observed.

Granitic Rocks

Granitic rocks form plutons that range in composition from quartz diorite to alaskite and intrude the rocks previously described. Each pluton has its own system of intrusive primary flow structures. Within this area, most plutons intrude older plutons rather than still older metamorphic rocks, and crosscut and truncate the flow structures of the older plutons.

The plutons will be described in order of their inferred ages. Field evidence indicates that the quartz monzonite of Mono Recesses is probably the youngest pluton but direct methods cannot relate the ages of the plutons southwest of it to those which are north and east of it. Studies in other areas have indicated broad groupings of plutons. The eastern plutons are in general younger, and within groups the age decreases with increase of alkali feldspar and quartz (Calkins, 1930; Krauskopf, 1953; Hamilton, 1956a; Evernden, Curtis, and Lipson, 1957). The assumption of such trends, coupled with the known field evidence for local sequences, suggests the approximate order of intrusion from oldest to youngest is: quartz diorite of Graveyard Lakes, Mt. Givens granodiorite, granodiorite of Shelf Lake, quartz diorite of Lee Lake, granodiorite of Chickenfoot Lake, granodiorite of Rock Creek, alaskite of Graveyard Peak, and quartz monzonite of Mono Recesses.

The granitic rocks form a series of transitional types sharing many features. Each pluton varies within itself in texture and composition, so that many plutons contain several rock types. Such variations have been studied quantitatively in other areas where the California batholith is exposed, with presentation by quartz-potassic feldspar-plagioclase diagrams (Rinehart and Ross, 1957), various mineral ratios (Chayes, 1956), and variation maps (Hamilton, 1956b). Mineral compositions (Table 1) vary irregularly with feldspar ratios. Rocks with high ratios of plagioclase to potassic feldspar are in general darker than those with low ratios and contain relatively more hornblende and less biotite, but such trends are irregular. Biotite is the sole mafic silicate in many of the leucocratic rocks. Plagioclase composition changes irregularly with rock type from calcic andesine in quartz diorite to albite or sodic oligoclase in alaskite.

With the conspicuous exception of the porphyritic quartz monzonite of Mono Recesses, most of the granitic rocks are equigranular, and the average grain sizes are between 2 and 5 mm. The rocks are commonly light or very light gray, and black mafic minerals speckle the generally near-white quartz and feldspars. Plagioclase more calcic than An_{20} is subhedral to euhedral in most rocks, zoned normally, in many places with oscillations (Pl. 4, fig. 3), and almost invariably has albite twins. In many cases it has Carlsbad and acline twins. Sodic plagioclase ($An < 20$) is generally anhedral and is zoned in some rocks but not others. Potassic feldspar is generally anhedral; the notable exception is in the euhedral phenocrysts, or perhaps more properly, porphyroblasts, of the quartz monzonite of Mono Recesses. Practically all thin sections show conclusive evidence of the partial replacement of plagioclase by potassic feldspar, with sodic rims on the plagioclase remnants; many have myrmekite along contacts. Quartz is light or very light gray in most rocks but smoky gray in alaskite. The common biotite is dark olive brown, but lighter-colored and greenish varieties are present locally. Hornblende is everywhere the common green variety.

All the granitic rocks have small amounts (commonly less than 1 per cent) of magnetite and sphene, less apatite, and traces of zircon and allanite. Red garnet is a rare accessory in alaskite.

The rocks are little altered. Most thin sections show a little chlorite (commonly altered from biotite), a little white mica in plagioclase cores, and perhaps a trace of epidote, but these secondary minerals do not, in most cases, exceed a fraction of 1 per cent. They will not generally be noted in the following pages.

QUARTZ DIORITE OF GRAVEYARD LAKES: Biotite-hornblende quartz diorite with unique texture crops out around Graveyard Lakes and is named informally for the lakes. (No specimens were collected east of the Lakes, and the extension of the pluton shown in that direction may include other rock types.) The rock is distingushed by its seriate subhedral plagioclase, which generally ranges from 0.2 to 5 mm but may be up to 10 mm in length. The plagioclase is calcic andesine, with normal-oscillatory zoning; the shapes are in general euhedral, but the margins conform in detail

to the surrounding grains. Plagioclase is light gray in hand specimen and thus darker than in most of the granitic rocks. About one-quarter of the rock is ragged poikilitic hornblende and biotite, the latter typically subordinate, in seriate crystals somewhat smaller than the plagioclase. The quartz is in small anhedra and in sprawling grains so extremely irregular that a single crystal may show in five separate areas in a thin section. The orthoclase or microcline is in small poikilitic anhedra, has a little myrmekite along plagioclase contacts, and is abundant enough locally so that the rock can be classed as a quartz-poor granodiorite.

In all thin sections, some of the large plagioclase crystals are seen to enclose sharply defined areas several millimeters in diameter with abundant tiny inclusions of biotite and hornblende. Some of these areas are in the cores of single plagioclase crystals, but others are shared by several large crystals. A microxenolithic origin for these tiny inclusion zones is suggested. Similar microxenoliths? were described by Bradley (*in* Bradley and Lyons, 1953, p. 104) in the Mt. Whitney area; from these he concluded that a metamorphic origin was indicated for the host granitic rock. This is unlikely.

Dark lenticular inclusions of all sizes from a fraction of an inch to tens of feet long, oriented parallel to primary flow structures, are common in the quartz diorite and locally crowd it densely. There are many areas of patchy migmatite in which abundant dark inclusions were veined and partially assimilated—streaked out into schlieren—by quartz diorite. Many of the inclusions are of fine-grained dark rocks, similar in appearance to the metadiorite to the north and possibly xenoliths of it; many others are of coarse amphibolitic rocks, some of them layered, composed of plagioclase and hornblende in different proportions, and containing green clinopyroxene in places.

MT. GIVENS GRANODIORITE: The bedrock of the southwest corner of the area belongs to the Mt. Givens granodiorite, named the Mt. Givens pluton by Hamilton (1956a), to the southwest[2]. Mapping by Krauskopf (1953), P. A. Lydon (unpublished report), and Hamilton shows that the pluton crops out in an area 10 miles wide and extends northwest (the boundaries are, as yet, unknown) for more than 30 miles. The rock is typically light-gray biotite-hornblende granodiorite, in which the hornblende is in distinct small prisms and is uniform over broad areas. Grain size is even, generally 2–3 mm. The plagioclase is subhedral and slightly zoned sodic andesine. Potassic feldspar forms late poikilitic anhedra; all feldspar is nearly white in hand specimen. Biotite and hornblende are in subequal amounts, totaling commonly about 15 per cent. Biotite is in thick ragged books, hornblende in slender prisms. Sphene is commonly euhedral.

Small dark inclusions of diverse textural types are displayed in most outcrops but are not abundant. The inclusions are finer-grained and richer in mafic minerals than the granodiorite and in general have a xenomorphic texture. Mineral proportions and textural details vary greatly between neighboring inclusions.

Obvious xenoliths are present locally. For example, beneath the Edison Lake high-water level near the upper end of the reservoir, there are numerous sharply bounded calc-silicate xenoliths in addition to the more common dark inclusions. The xenoliths have maximum dimensions of a few feet and parallel the dark inclusions. Many of the xenoliths show marginal transformation to amphibolite. One such xenolith, exposed as a rectangular block 3 by 6 feet, shows marginal amphibolitization and the spalling off of fragments of the resulting amphibolite (Pl. 5, fig. 4). The hornfels is fine-grained and contains more than 50 per cent plagioclase and different proportions of green clinopyroxene, hornblende that is green in both specimen and section, quartz, and brown garnet. Within a fraction of an inch of the granodiorite contact, the small anhedra of pyroxene and green hornblende were replaced by large poikilitic sieves of hornblende, black in hand specimen, with inclusions of quartz and feldspar (Pl. 4, fig. 6). Some of this black hornblende reacted in turn to form biotite and sphene. Thin sections show that some of the poikiloblastic hornblende was rafted as xenocrysts into the granodiorite, and that some remained in discrete dark inclusions. P. A. Lydon has found a pendant containing metasedimentary rocks less than 3 miles south of this location. The xenoliths presumably came from the northward extension, since removed by erosion, of that mass.

It is inferred that many of the other dark inclusions in the granodiorite similarly originated as hybridized xenoliths; proof of such

[2] In plutonic rock names, "pluton" is considered comparable to "formation" in sedimentary rocks. Where a pluton is dominantly of one rock type, that type is properly used in the name, as "Mt. Givens granodiorite." Where several rock types are abundant, pluton is better used, as Mt. Givens pluton. The parallelism with stratigraphic terminology is obvious.

an origin for many more inclusions was found in the granodiorite of Rock Creek and is discussed with the description of that pluton. Contamination of the Mt. Givens granodiorite by pyroxene hornfels was documented to the southwest by Hamilton (1956a), and the conclusion reached that most of the mafic mineral content of the pluton might be due to assimilation of more mafic wall rocks.

Dark dikes a few inches to a few feet thick with sharp contacts and parallel walls cut the granodiorite in many places. The dike rocks are mesocratic diorites and quartz diorites, commonly fine-grained and strewn with larger crystals of plagioclase. Average grain size is about 0.2 mm. Plagioclase (andesine) occurs mostly as small anhedra but also as large subhedra and is zoned normally in some specimens but is unzoned in others. Small amounts of potassic feldspar occur as small interstitial untwinned grains, and small quartz anhedra make up 5–10 per cent of the dike rocks. Hornblende is variously in anhedra, subhedral prisms, and euhedral needles, and hornblende and biotite total 10–25 per cent of the rocks. Many dark dikes have a faint planar orientation of mineral grains parallel to their walls, and some dikes have thin borders darker than their interiors. Pegmatite dikes commonly cut the dark dikes where both occur together.

GRANODIORITE OF SHELF LAKE: Biotite-hornblende granodiorite distinguished by very large prisms of hornblende crops out in two masses in the southwest, and the rock of the two areas may have been continuous before intrusion of the intervening alaskite of Graveyard Peak. The pluton is named informally for Shelf Lake, within its western area.

About 20 per cent of most specimens is biotite and hornblende, the latter generally dominant. The maximum length of the well-shaped stubby prisms of hornblende ranges from 5 to 15 mm in different parts of the pluton; uncommonly, cores of augite lie in the hornblende. Biotite is also much larger than in most other plutons, commonly as large as 5 mm. The black mafic minerals stand out strikingly against the nearly white quartz and feldspar. Plagioclase is mostly in subhedra of calcic andesine 1–5 mm long, and orthoclase occurs as irregular poikilitic crystals that partly replace plagioclase.

Lenticular dark inclusions are common to abundant, and steep primary flow structures are well developed in much of the pluton. In the western area there is much patchy migmatite, in which abundant large dark inclusions and long feathery masses show all stages of marginal assimilation. Both the granodiorite and the dark inclusions in and near such migmatite vary widely in texture and composition. The geometry of these diversely oriented inclusions shows diking, riving, and rafting away by the granodiorite and specifically does not show replacement in place except on a tiny scale.

QUARTZ DIORITE OF LEE LAKE: A pluton of uncommonly calcic quartz diorite extends into the north edge of the area, where it is in contact only with metavolcanic rocks. Plagioclase is light gray, so the rock looks darker than its mafic-mineral content of 10 per cent would suggest; biotite is alone or dominant over hornblende, and the common small size of both gives a further apparent darkness. Plagioclase is subhedral; many grains in some specimens have extremely calcic (An_{85}) corroded cores, sharply separated from the enveloping, slightly zoned calcic andesine, which in turn gives way smoothly outward to markedly graded rims with marginal compositions near An_{25}.

Near its southwest contact with metavolcanic rocks, the quartz diorite is greatly sheared parallel to the contact and to the shearing in the metatuff, and a flaser gneiss is developed locally. Away from this part of the contact, quartz is greatly strained, but no other effects of cataclasis except a little bent biotite were noted. The cataclasis may be due to the forceful intrusion of the quartz monzonite of Mono Recesses only 200 yards distant, as the marginal shearing in and near that pluton along the metavolcanic rocks was extreme.

The actual contact between quartz diorite and metatuff is sharp even in thin section. The metatuff is greatly sheared and carries augen of microcline that might be either porphyroblasts produced by contact metamorphism or recrystallized relics from the original rocks. The fine biotite in the metatuff is partly brown, partly green; grains of the two colors are either strewn together or crudely segregated; biotite in the quartz diorite is brown, and many grains have green tips.

Modal analyses by C. D. Rinehart and D. C. Ross (personal communication) of specimens from this pluton in the quadrangle to the north show a similar content of mafic minerals, but enough potassic feldspar—12–30 per cent—to qualify the rock as a granodiorite, and only normal plagioclase, zoned An_{45}–An_{25}, without calcic cores.

GRANODIORITE OF CHICKENFOOT LAKE: The southeast corner of the area is underlain by a pluton of hornblende-biotite granodiorite with subordinate quartz monzonite. Feldspars are

2–10 mm long, and both plagioclase (zoned sodic andesine) and microcline are light gray. Biotite and hornblende, either of which may be dominant, total 5–15 per cent and are mostly in clusters of small poikilitic anhedra. Biotite is brown in some specimens, green in others. Quartz is extremely irregular, and microcline is anhedral and poikilitic.

Dark inclusions are abundant locally. Flow structures are in general weak or lacking. At the contact with metagabbro at Long Lake, a foot or so of patchy mixed rock separates massive granodiorite of superficially normal appearance from metagabbro, and inclusions (xenoliths of metagabbro?) are small and few. This marginal granodiorite is unusual in thin section, however, as it contains about 10 per cent of myrmekite and quartz-microcline micropegmatite (Pl. 6, figs. 3 and 4); also, its hornblende and biotite are ragged and poikilitic, and much of its quartz is in irregular aggregates of diversely oriented anhedra. The hornblende and biotite are strikingly similar to those in the near-by metagabbro in texture, size, and appearance, and a xenocrystic origin for them is likely. This contaminated granodiorite contains 10 per cent biotite and hornblende, of which practically all looks xenocrystic, yet its bulk composition is similar to that of the rest of the pluton; the gabbro contains about 40 per cent hornblende and biotite. Assimilation of one part of metagabbro in three parts of a silicic magma is suggested as a possible reaction that gave the pluton its final composition. If average compositions of 60 per cent of plagioclase An_{55} and 40 per cent hornblende and biotite are assumed for the metagabbro, and 50 per cent of An_{35}, 10 per cent hornblende and biotite, 20 per cent potassic feldspar, and 20 per cent quartz for the granodiorite, then calculation indicates a hypothetical initial magma, without mafic content, with a composition equivalent to 46 per cent An_{37}, 27 per cent orthoclase, and 27 per cent quartz. These figures make no allowance for variations in mafic mineral compositions, nor for such changes as the addition of potassium and subtraction of silicon needed to convert hornblende to biotite. Assimilation of such magnitude may have been a major factor in producing variations in some of the granitic rocks of the California batholith (Hamilton, in press).

GRANODIORITE OF ROCK CREEK: The northeast margins of the area are underlain by a pluton of hornblende-biotite granodiorite and subordinate quartz monzonite and quartz diorite. The pluton is separated within the quadrangle by the quartz monzonite of Mono Recesses into two areas, but mapping by P. C. Bateman shows that the quartz monzonite ends a short distance to the east. The granodiorite has 5–20 per cent, but for the most part about 10 per cent, of biotite and hornblende; the former in most cases is slightly dominant. The rocks are equigranular; grains of all minerals are about 2–3 mm in diameter, and the mafic grains are distinct rather than in aggregates. Plagioclase is subhedral sodic andesine showing normal oscillatory zoning. The potassic feldspar is anhedral and poikilitic; some shows grid twinning, and some is microperthitic. Feldspars are white in hand specimen.

The granodiorite intrudes metagabbro and metasedimentary rocks and is in turn cut by the quartz monzonite of Mono Recesses. The contact at Box Lake with the eastern metagabbro is marked by abundant xenoliths (Pl. 5, fig. 1) in a matrix of sugary granodiorite. In thin section, this contact granodiorite is seen to contain abundant small fine-grained xenolithic clusters and larger xenocrysts of poikilitic hornblende and biotite. Some of the plagioclase, otherwise about An_{40}, has irregular cores of xenocrystic labradorite. The larger "granitic" quartz and feldspar crystals are anhedral. Despite the odd texture and the obvious evidence of assimilation, the composition of the granodiorite is about typical of the pluton. As do the comparable relationships of the granodiorite of Chickenfoot Lake to the south, these features suggest major contamination of an initially more silicic and sodic magma by mafic rock, with thorough mixing throughout the pluton of the hybrid materials.

The contact with metasedimentary rocks in the north-central part of the area is well exposed. At the extreme north edge of the quadrangle, granodiorite lies parallel to and east of the steeply dipping metasedimentary rocks on the upper slopes, but at lower levels the granodiorite underlies the metamorphic rocks with a nearly horizontal contact and truncates their isoclincal structures at right angles. Most of the rest of the exposed contact dips under the metasedimentary at moderate angles, and almost everywhere the metamorphic rocks are sharply truncated. Around Big McGee Lake, blocky xenoliths from a few inches to several hundred feet across are abundant, and the lake probably lies not far below the former undulating roof of that part of the pluton. The truncation of the metamorphic rocks against such a roof and the abundant large

xenoliths near the roof show that stoping was here a major process of emplacement. Xenoliths other than the small dark inclusions are virtually nonexistent in the rest of the exposed part of the pluton, so most stoped blocks must have been removed either by sinking or by assimilation, or both.

Dark inclusions, most of which are only a few inches long, are common in much of the granodiorite and are typically a few yards apart on outcrops. Such dark inclusions were described by Pabst (1928), who found little to suggest an origin for them and concluded that they were probably "autoliths". The present area presents proof that some such inclusions are produced by the hybridization of calc-silicate xenoliths. Dense swarms of xenoliths, which progressively change in character away from the contact, lie at the east contact of the metasedimentary rocks, here calc-silicate rocks (locally amphibolitized), produced by contact metasomatism of limestone. At the contact, the xenoliths, which are of uniform size, are of calc-silicate rocks with narrow amphibolitized margins (Pl. 5, fig. 2). Even within a foot of the main sharp contact, the amphibolitic rims have much of the typical xenomorphic texture of the dark inclusions of puzzling origin in the pluton interior and have such typical features as porphyroblasts of potassic feldspar shared by both inclusion and host granodiorite (Pl. 6, figs. 5 and 6). In the granodiorite, the gray amphibolitized rims of the xenoliths increase in thickness inward at the expense of the green calc-silicate assemblages, and the amphibolites are in turn progressively altered to the compositions — mafic biotite-hornblende-andesine quartz diorite and granodiorite—typical of the dark inclusions of the pluton interior. Only the near-contact inclusions retain a weak and contorted schistosity, mimicked by the new and enlarged biotite and hornblende as shown by the photomicrographs. The grain size in inclusions a foot from the contact is in general near 0.3 mm. Thirty feet away, the grain size is 0.5 mm, biotite is commonly dominant over hornblende, and calc-silicate minerals are absent in most inclusions. However, the inclusions retain much of the angular form of the initial xenoliths and are oriented by flow of the granodiorite (Pl. 5, fig. 3). Still farther from the contact, the inclusions, sparser owing to dilution by granodiorite, have the typical flow-softened lines of the inclusions farther within the pluton. These progressive changes are recorded in all stages of completion, and individual inclusions illustrating each stage overlap into zones of the xenolith swarms dominated by inclusions representing other stages.

Numerous large xenoliths of metamorphic and hydridized rocks as much as several hundred feet in diameter lie near the McGee Creek trail a few hundred yards northeast of Big McGee Lake. These rocks are cut by thick dikes of granodiorite and are commonly superficially uniform fine-grained biotite-hornblende-plagioclase rocks, probably hybridized calc-silicate xenoliths. Locally present in these dark-gray rocks are small green spheres of poikilitic epidote with a little quartz and plagioclase in ellipsoids of leucocratic fine-grained orthoclase-quartz-hornblende-clinozoisite-plagioclase rock (Pl. 2, fig. 3).

ALASKITE OF GRAVEYARD PEAK: An alaskite pluton with facies of biotite granite is the youngest pluton southwest of the quartz monzonite of Mono Recesses and is here named informally for the highest peak composed of it. A remarkable stoping-assimilation complex, described on later pages, is developed between the alaskite and adjacent mafic rocks.

The typical composition is about 30 per cent quartz, 50 per cent perthitic microcline, and 20 per cent sodic feldspar (about An_{10}), and less than 1 per cent of biotite and minor accessory minerals. The feldspars are cream and the quartz is smoky gray in alaskite; the feldspar is white, the quartz light gray, in biotite granite. Quartz and feldspars are anhedral, in many cases extremely so, and sodic feldspars in most cases show little zoning. Potassic feldspar is commonly coarsely perthitic microcline (Pl. 4, fig. 5). Grain size varies widely, but the common average is 3–4 mm; in many places the average grain size is 1 cm. Elsewhere the rock is relatively fine-grained.

Within the eastern part of the alaskite mass are several small masses of fine-grained biotite quartz monzonite. Their relationship to the alaskite is not known, and they are not separated from it on the geologic map. Other small masses, also not shown, of quartz diorite, biotite-hornblende diorite, and amphibolite are presumably xenoliths like those near Graveyard Peak.

QUARTZ MONZONITE OF MONO RECESSES: Underlying more than half the area is a pluton of coarse-grained and commonly porphyritic leucocratic quartz monzonite with granite facies. This is the central and widest part of the pluton, which Mayo (1941) found to have a northwest length of 30 miles. The pluton seems

lithologically identical to the Cathedral Peak granite of the Yosemite region (Calkins, 1930), but the two plutons are not continuous, and a different name (derived from the deep tributary valleys of Mono Creek, the four Recesses) is used here. Rocks of the pluton are massive and particularly resistant to weathering and form most of the higher peaks. Both porphyritic and equigranular rocks occur in all parts of the pluton, but porphyritic types are dominant. The average grain size ranges from 3 to 10 or, uncommonly, 15 mm. Mafic minerals, biotite alone or dominant over hornblende, total less than 5 per cent.

Microcline is in general dominant over the zoned calcic oligoclase and is white or very pale orange or pink. In most of the pluton, it forms both anhedra and large euhedral phenocrysts. Phenocrysts are rectangular, in many areas 2 inches long, and occur in all proportions from uncommon and scattered to densely crowded (Pl. 2, fig. 2). Both anhedra and large euhedra of microcline contain small inclusions of all other minerals. Near-by small inclusions of plagioclase are in many cases optically continuous with narrow sodic margins along the contacts, which in many cases are discordant to the zoning of the plagioclase and are thus shown to be remnants of larger crystals almost entirely replaced by microcline (Pl. 4, fig. 4); all stages of such replacement are recorded. Phenocrysts are typically non-oriented; even in some of the places where they are more or less parallel, formation of the phenocrysts by replacement in a stressed but nearly static environment rather than orientation by magmatic flow seems indicated (Pl. 3, fig. 4). The phenocrysts certainly formed late in the crystallization of the pluton and are perhaps products of the crystallization of and reaction by deuteric liquids prior to complete solidification. The large crystals might be termed porphyroblasts or megacrysts, but the familiar phenocrysts is used here with a geometric rather than genetic meaning. Many of the phenocrysts are zoned concentrically, and amounts of included minerals vary from ring to ring; concentric growth with varying "replacement energy" seems indicated.

Plagioclase is white and typically subhedral and has large oscillatory-zoned cores and rapidly graded rims. Forty-two reversals in zoning were counted in a typical crystal. Most quartz is in large irregular grains, but in most specimens some is in ellipsoidal aggegates of sutured anhedra. Biotite and hornblende grains are in general smaller than 2 mm and commonly are found in small clusters.

The quartz monzonite characteristically contains sparse schlieren but only very rare dark inclusions. All other plutons except part of the alaskite carry dark inclusions in much greater abundance, and no others carry schlieren, except in small areas where contamination by assimilation is probable.

Several small masses of diverse quartz monzonites, at least one of which is bounded sharply against the surrounding porphyritic quartz monzonite, are present near Helen Lake in the northwest part of the area but were not mapped separately. The rocks are characterized by ellipsoidal quartz aggregates. The most abundant types are fine-grained biotite quartz monzonite and granite that contain scattered poikilitic sieves of potassic feldspar as much as 1 cm long; coarser-grained biotite quartz monzonite is less common. The only contact of such rocks seen was at an altitude of 9950 feet by the Muir trail. It shows 6 inches of smoothly gradational textural change along an irregular contact between normal porphyritic quartz monzonite and fine-grained seriate quartz monzonite.

Contacts of the pluton are in many places characterized by large and extremely abundant dikes sent out into the wall rocks. These injection complexes and the protoclastic border zone of the pluton against metavolcanic rocks are described on later pages.

COMPLEX OF GRAVEYARD PEAK AREA: A stoping-assimilation complex in which the alaskite of Graveyard Peak engulfed and assimilated metadiorite and the quartz diorite of Graveyard Lakes is superbly exposed about Graveyard Peak and Lake Marilyn. Most contacts dip under the dark rocks, which presumably are truncated completely at depth by the alaskite. On the east side of Graveyard Peak, the contact is irregular, flow structures in the dark rocks are cut at all angles, large xenoliths are enclosed in alaskite, and dikes from the alaskite cut the dark rocks (Pl. 7, fig. 1). Irregular intrusive masses of alaskite lie within the quartz diorite and enclose many xenoliths of the quartz diorite.

In the high basin 1 mile northeast of Graveyard Peak, a layered complex concordantly underlies the gently dipping contact between alaskite and metadiorite. Abundant xenoliths of metadiorite an inch to several hundred yards across are strewn through the alaskite (Fig. 1). Their contacts show assimilation by alaskite, and dark xenocrysts decrease in abundance in the alaskite away from the xenoliths (Pl. 7, fig. 3). In both xenolith swarms in the alaskite and in small irregular intrusions

of alaskite in metadiorite, patchy assimilation-migmatites are widespread and contain metadiorite inclusions in all stages of digestion in alaskite and in hybrid quartz monzonite formed by contamination. The alaskite within several hundred feet of the contact with the overlying metadiorite is massive and coarse-grained, but beneath this is a 500-foot-thick complex of layered alaskite with graded content of xenocrysts. Individual layers are a few inches to a few feet thick. Almost every layer is graded, containing abundant dark minerals at the base in sharp contact with the underlying layer and becoming gradually lighter upward (Pl. 7, fig. 2); the upper halves of many layers contain no dark minerals. The layers dip about 30° south and obviously record a magmatic sedimentation, whereby the dark grains sank within each layer. Numerous low-angle "unconformities" suggest that each layer formed as a discrete internally flowing unit rather than as a static layer. Xenoliths in the layered complex are tabular and concordant, in contrast to the blocky or complexly irregular xenoliths in the massive alaskite above (Fig. 2). The weak foliation and lineation of the metadiorite in both types of xenoliths is truncated at all angles, and the xenoliths have been diversely rotated.

Farther down in the layered rocks, coarse-grained quartz monzonite, both with and without large phenocrysts of potassic feldspar, forms many layers and groups of layers intercalated with alaskite layers. Quartz monzonite increases in abundance downward, and the contact between the layered complex and the main mass of quartz monzonite of Mono Recesses to the north is gradational. Some quartz monzonite is presumably a product of contamination of alaskite, but much is apparently similar to the main mass below and to the north. The interlayering of quartz monzonite and alaskite indicates that both were mobile at the same time, and other relationships show that each was solid in some places while the other was still mobile. Thus a xenolith of alaskite containing a xenolith of metadiorite was noted in porphyritic quartz monzonite. Near by, a large xenolith of metadiorite is cut by a dike of porphyritic quartz monzonite, and both are cut by an alaskite dike. Both dikes are about 15 feet thick and are sharply bounded against metadiorite, but the quartz monzonite-alaskite contact is irregular and gradational over several inches. Another near-by outcrop displays several graded layers in contaminated alaskite overlying and truncating the flow structure of porphyritic quartz monzonite (Pl. 7, fig. 2). This lower rock gives way through a hidden contact some tens of feet below to more graded layers.

Detailed age relationships are thus ambiguous and contradictory and show that the solidification of the alaskite and the quartz monzonite of Mono Recesses was essentially contemporaneous near their contact. However, because

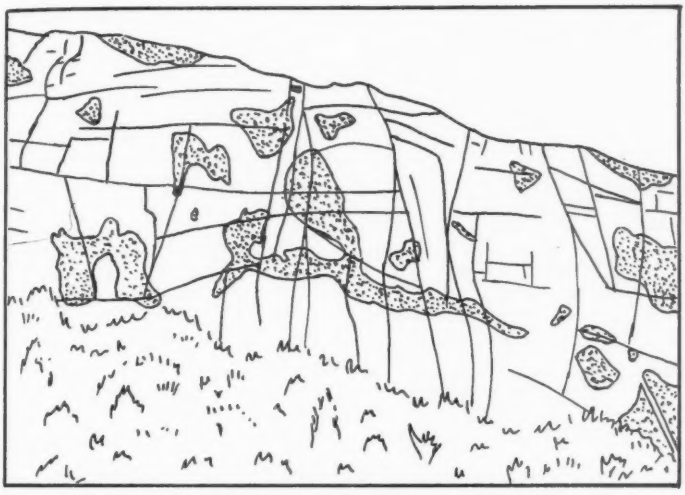


FIGURE 1.—XENOLITHS OF METADIORITE IN ALASKITE
The jointed cliff is 50 feet high. Complex of Graveyard Peak 0.1 mile east of south end of Marilyn Lake

the layered complex gives way downward to quartz monzonite, which forms a far larger pluton than the alaskite, it is inferred that the bulk of the quartz monzonite probably crystallized after the alaskite. In the coarsely porphyritic quartz monzonite beneath the layered rocks, flow structures are broadly concordant. Schlieren are abundant near the top but rare below a few hundred feet beneath it.

The west part of the metadiorite-quartz diorite-alaskite complex is almost comletely exposed, but only its more prominent features have been noted in the present study.

CONTACTS OF QUARTZ MONZONITE OF MONO CREEK: Along much of its contact, the quartz monzonite sent out great quantities of dikes. The dikes are of aplite (commonly with internal pegmatite), of pegmatite alone, and of varied quartz monzonites, both equigranular and porphyritic. Most dikes dip gently to moderately toward the quartz monzonite and thin from maximum thicknesses of 100 or even 200 feet and decrease in number away from the contacts. The farthest-traveled dikes reach a mile or more into the walls. Near the contacts the dikes branch and join.

In the broad injection complex between the quartz monzonite of Mono Recesses and the adjacent masses in the southeast corner of the area, dikes are extremely thick and abundant (Fig. 2) and are injected into a wide variety of host rocks. The anastomosing dikes are subparallel and have split the host rocks into sharp-pointed lenses and produced a layered structure. Size and abundance of the dikes decrease eastward, and the east margin of the complex is designated arbitrarily. Near the west margin, fully half the height of most of the cliffs on which the complex is exposed is of dike rocks. The varied host rocks of the complex include the granodiorites of Rock Creek and Chickenfoot Lake, the eastern metagabbro and many smaller masses of similar metagabbro, and mafic metamorphic rocks which represent a discontinuous screen between the quartz monzonite and the granodiorites. Near the east end of Ruby Lake, dikes make up three-fourths or more of the complex, which

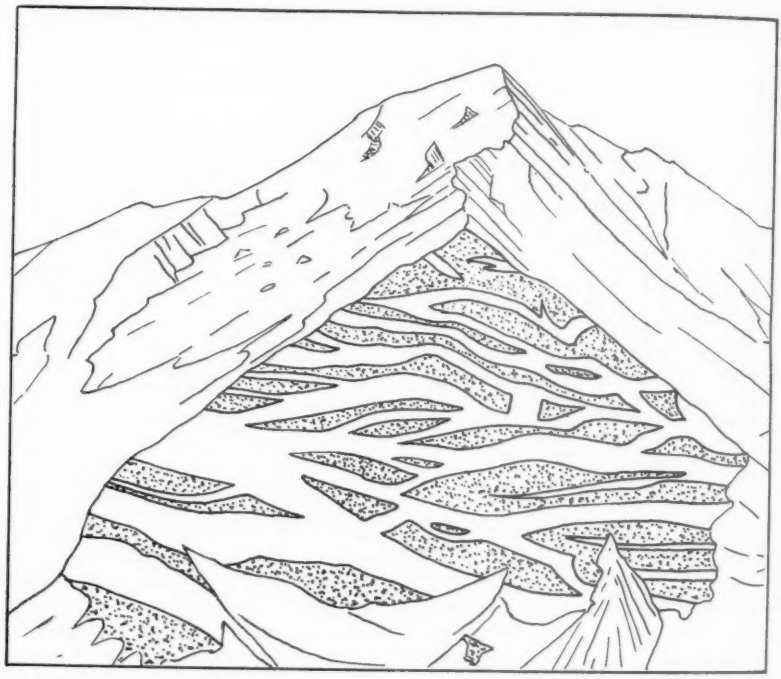


FIGURE 2.—INJECTION COMPLEX OF DIKES FROM QUARTZ MONZONITE OF MONO RECESSES INTO MAFIC ROCKS

The 800-foot-high diked cliff is 0.4 mile west of Dade Lake and appears just left of center in Figure 1 of Plate 2.

there becomes a layered series of large xenolith lenses, the largest 200 by 20 feet, of amphibolite. Contacts are sharp, and swarms of small xenoliths are present locally.

Similarly, abundant dikes were injected into the metavolcanic rocks along much of their contact with the quartz monzonite, again producing a dike-layered complex in places. However, the most striking feature along this contact is the strong protoclastic foliation in the quartz monzonite. The intensity of cataclasis decreases into the pluton, but most of the quartz monzonite within half a mile of the contact is strongly sheared. Poorly defined zones of flaser gneiss, in which finely crushed quartz, feldspar, and most of the mafic minerals form folia between augen or laminae of flattened quartz and partly granulated feldspar (Pl. 6, fig. 1), alternate with zones of mortar gneiss. The mortar gneiss, although it may look superficially massive and uncrushed in hand specimen, is seen in thin section to have granulated margins about most grains. The crushed rocks have partially recrystallized and are hard and solid. Cataclastic effects are also obvious in the metatuffs, although in them it is not possible to distinguish between the shearing due to pre-intrusive regional metamorphism and that due to movement of the quartz monzonite. The dikes in the metamorphic rocks, however, were greatly sheared and converted to flaser or mortar gneisses (Pl. 6, fig. 2). Some of the dikes were sheared so greatly that they intertongue cataclastically with the comparably sheared metatuffs. That the cataclasis was protoclastic, caused by the dragging upward of the solid margin of the pluton by the mobile interior, is shown by its restriction and parallelism to this one contact, and by its giving way inwardly to normal flow structures. Such a contact is rare in the Sierra Nevada, although D. C. Ross (personal communication) states that the contact of the porphyritic quartz monzonite and metavolcanic rocks in the quadrangle to the north is similar. Along Mono Creek between the First and Second Recesses, slightly crushed porphyritic quartz monzonite with a weak cataclastic foliation is present, probably indicating minor interior protoclasis.

The profusely diked eastern contacts of the quartz monzonite within the area are as far as known restricted to the north and south ends, which are crudely in line but separated by the big northeastward "embayment" of the pluton. It is an obvious inference that the contact was once continuous across this "bay", but that at a late stage of development the pluton broke through this contact and established the final contact 5 miles to the northeast. Mayo (1941, p. 1037) found that this bulge in the pluton is followed concentrically to the northeast by structural trends in both metamorphic and granitic rocks.

The contact between the granodiorite of Shelf Lake and the quartz monzonite of Mono Recesses is exposed throughout most of its length. Within several hundred yards of the contact, dikes as much as 20 feet thick of pink aplite with irregular interior facies of pegmatite cut the granodiorite. Most of the larger dikes dip gently. Within 100 yards of the contact, dikes lace tightly through the granodiorite; many of these dikes are of quartz monzonite, either porphyritic, like the main pluton, or equigranular. Xenoliths, many of them partly digested to schlieren, are common in the dikes. Most dikes stop at the quartz monzonite contact, but some continue into it. The contact is smooth at map scale but irregular in detail and sharply truncates the flow structures in the granodiorite. Actual contacts are sharp; any gradation or marginal reaction is conspicuous only in a zone an inch or so wide. The few xenoliths of granodiorite in the main quartz monzonite near the contact have fuzzy outlines and are sprinkled with large euhedral porphyroblasts of potassic feldspar, like the phenocrysts of the quartz monzonite. The uncommon schlieren in the quartz monzonite parallel the steep contact, but phenocrysts are not oriented.

The contact between the granodiorite of Rock Creek and the quartz monzonite of Mono Recesses is sharp in the east but apparently gradational in the west. Above Ruby Lake, the weak flow structure of the granodiorite is truncated against the contact with phenocryst-crowded quartz monzonite (Pl. 2, fig. 2). The quartz monzonite grades into pegmatite an inch or so thick along most of the contact. This pegmatite is in sharp contact with the granodiorite. The weak flow structure of the quartz monzonite, shown by the few schlieren in it, is subparallel to the contact. Near Hilton Creek, D. C. Ross (personal communication) found the sharp contact exposed, with flow structures in the granodiorite again truncated against the contact. Ross reports that light-colored dikes as much as 50 feet thick, possibly offshoots from the quartz monzonite, are common in the cirque west of Davis Lake.

In Pioneer Basin, the contact seems to be gradational over about two-thirds of a mile, and porphyritic quartz monzonite and equi-

granular granodiorite are both present in vague irregular zones. No sharp contact was observed between the southernmost structural observations shown in Pioneer Basin and the north edge of the map. If any sharp contact is present in this part of the area, it must lie north of Mono and Golden Creeks but south of Pioneer Basin. Fresh rock is continuously exposed in Pioneer Basin, and all lithologic changes in it were investigated carefully. In the basin south of the recorded flow structure dipping 85° north (Pl. 1), the rock is uniform porphyritic quartz monzonite, with, as is characteristic of the Mono Recesses pluton, sparse schlieren and extremely rare dark inclusions; phenocrysts are ½- by 1-inch rectangular euhedra strewn without orientation in a coarse groundmass. Near the 35-degree-dip flow structure, the quartz monzonite changes progressively over many tens of feet and becomes obscurely porphyritic northward where its phenocrysts are less well shaped and smaller, but the groundmass stays the same. The uncommon small dark inclusions become more numerous. Within the next several hundred yards northward, porphyritic and nonporphyritic facies are irregularly intermingled, and no actual contacts are present; the apparent differences are only in the size, abundance, and shape of phenocrysts. The porphyritic rocks decrease in abundance northward, and none are present north of the 60-degree dip symbol on Plate 1. The inferred contact is drawn through a faint contact, smoothly gradational over 5–10 feet, separating similar rocks. To the south is quartz monzonite with thin books of biotite; to the north is granodiorite with biotite in thick books. Dark inclusions are more common to the north. Still farther north, the granodiorite changes smoothly over tens of yards to quartz monzonite which is indistinguishable in hand specimen from the nonporphyritic rock to the south. The available evidence suggests that the quartz monzonite of Mono Recesses and the granodiorite of Rock Creek were mobile simultaneously and were physically mixed in a broad contact zone. If, on the other hand, a contact was overlooked between Pioneer Basin and Mono Creek, then the granodiorite of Rock Creek is represented in Pioneer Basin by a facies of porphyritic quartz monzonite closely similar to that of the Mono Recesses pluton.

OTHER CONTACT RELATIONSHIPS: The quartz diorite of Graveyard Lakes is inferred to be younger than the adjacent metadiorite because of its nonmetamorphic character and the possible xenoliths of metadiorite within it.

The only contact of the Mt. Givens granodiorite seen is on the north side of the outcrop band of the granodiorite of Shelf Lake, where the contact is smoothly gradational over 1–2 feet and is obscure because of the similarity of the two granodiorites. The Mt. Givens is massive, has less hornblende and biotite than the Shelf Lake, and lacks the big hornblendes that characterize the Shelf Lake. Dark inclusions are common in the Shelf Lake, uncommon in the Mt. Givens. The apparently dikelike form and concordant flow structures of the Shelf Lake suggest that it is the younger. Both granodiorites here are cut by abundant dikes of aplite and pegmatite within a few hundred yards of the quartz monzonite of Mono Recesses.

A few small xenoliths of the distinctive granodiorite of Shelf Lake were seen near the southern contact of the western mass of alaskite. Many dikes and small irregular intrusions of aplite and aplitic biotite granite cut the granodiorite near by. Near Mono Creek west of its North Fork, abundant dikes up to 30 feet thick of aplite (in many places with internal facies of biotite granite) and of coarse biotite leucogranite cut the granodiorite and contain uncommon xenoliths of it.

STRUCTURES IN GRANITIC ROCKS: Most of the granitic rocks contain primary flow structures of the types described from the Sierra Nevada by Cloos (1936), Balk (1937), and Mayo (1941). Several of Mayo's illustrations are of structures within this area: his figure 10*C* and figure 1 of Plate 8 illustrate flow structures, and his Plate 12 illustrates joints. Flow structures are recorded by the dimensional orientation of xenoliths and dark inclusions and to a far less obvious extent by the orientation of mineral grains. These structures record motion at times when the plutons contained enough liquid to flow without fracture of grains yet were solidified enough to retain mineral orientations. In the quartz monzonite of Mono Recesses, flow structures are recorded by schlieren, locally parallel, elsewhere swirled. Such schlieren are in some places discordant to the pervasive primary flow structures, so that a later shearing before the final consolidation took place is indicated, as Mayo (1941, p. 1028–1029) emphasized. The late structures may be related to the protoclastic border zone.

Structural observations are too few to permit drawing satisfactory flow-trend lines on the map, in the manner of Cloos and Mayo, or to evaluate Mayo's schematic map (Mayo, 1941, Pl. 1). An arch is perhaps suggested in the

northwest part of the quartz monzonite of Mono Recesses. The only generalization that can be made about the flow structures elsewhere is that they are commonly steep with a down-dip lineation and parallel contacts with older rocks near those contacts.

The granitic rocks are jointed in diverse directions. Parallel joints are commonly a few feet to a few tens of feet apart. In small outcrops the diversity of directions is confusing, but in large outcrops, or on air photographs, two conspicuous sets of crudely perpendicular steep joints can generally be distinguished from the maze of joints in other directions. Throughout most of the area, the strongly dominant joints strike northeastward, crossing pluton contacts indiscriminately. Conspicuous joint sets related to flow structures or to contacts of single plutons were noted only in the unique protoclastic border of the quartz monzonite of Mono Recesses. Obviously at least the major joints represent strain due to regional stress rather than to stresses unique to each pluton.

MECHANISM OF INTRUSION: All the granitic rocks formed as intrusive plutons which are discordant to older structures in both metamorphic and granitic rocks. They are broadly uniform and in general have only slight changes toward diverse wall rocks. Internal flow structures prove mobility and flow within each pluton. Contact metamorphism was superimposed on the low-grade regional metamorphism of the old sedimentary and volcanic rocks. The low grade of the regional metamorphism suggests that the plutons were intruded at a depth of not more than a few miles. The plutons were intruded as mobile masses discordant to their walls structurally, physically, and chemically.

Evidence for a variety of mechanisms has been presented on prior pages. The vertical and lateral advance of some plutons by large-scale stoping is seen in stoping of metasedimentary rocks by the granodiorite of Rock Creek, stoping of metadiorite and quartz diorite by the alaskite of Graveyard Peak, and lens-stoping of varied older rocks by the quartz monzonite of Mono Recesses. Smaller-scale stoping is demonstrated by the granodiorite of Rock Creek against the eastern mass of metagabbro. Stoping of huge blocks, with dimensions perhaps of miles, might be expected in a batholithic complex such as this, but no evidence for it was recognized within the area. If most of the dark inclusions are xenoliths, they would also indicate extensive stoping; however, many of these inclusions may represent fragments from levels perhaps as deep as those of the formation of the magmas themselves.

Blocks stoped by the granodiorite of Rock Creek may have been largely assimilated, and the ferromagnesian and calcic components of this rock and of the granodiorite of Chickenfoot Lake can be explained in terms of the assimilation of one part of mafic rocks to three parts of leucogranodiorite magma. The dark inclusions of the granodiorite of Rock Creek formed at least partly, and conceivably entirely, from the hybridization of xenoliths. As such inclusions are strewn throughout the pluton, apparently by flow circulation, a nearly complete mixing of assimilation products might be expected. The calcic cores of many plagioclases in the quartz diorite of Lee Lake suggest assimilation of calcic rocks, but the known wall rocks of that pluton are too sodic to have yielded such possible xenocrysts. Quartz monzonite was formed locally by assimilation of mafic rocks in the alaskite of Graveyard Peak, but the felsic composition of that mass shows that assimilation of mafic and calcic rocks could have contributed little to its exposed bulk. Neither could the leucocratic quartz monzonite of Mono Recesses have assimilated large quantities of more mafic rocks, and if large volumes of blocks were stoped into it, they must have sunk out of sight.

Evidence for forceful injection is given by the protoclastic border of the quartz monzonite of Mono Recesses against metavolcanic rocks and by the dike swarms along some contacts of that pluton. The marginal dikes, dipping toward the pluton and representing in places more than half the height of the exposed contact, indicate lifting, perhaps accompanied by marginal thrusting, of the older rocks. The contacts seen of the other plutons are unsheared, and there is no suggestion of great force exerted by plutons on their walls. The exception, the crushed rocks of the west margin of the quartz diorite of Lee Lake, are best interpreted as due to shearing by the quartz monzonite of Mono Recesses, only one-eighth of a mile away. The marginal flow structures of the various plutons prove motion but indicate nothing about force applied beyond their margins. The extent to which deformation of the metamorphic rocks is due to intrusive plutons is conjectural. Intrusive plutons, however, cannot have caused isoclinal deformation of roof rocks. Noble (1952), arguing largely by analogy with small intrusions in the Black Hills, concluded that forceful intrusion of Sierra plutons might leave little conclusive evidence for its operation against

schistose wall rocks; but the theory of forceful injection as the dominant mechanism of intrusion in the Sierra Nevada fails to explain the roof-rock deformation, the abundant evidence of passive processes at many contacts, and the intrusion of granitic plutons into older plutons without screens of metamorphic rocks.

Mayo recognized the problems besetting the theory of forceful injection and concluded (1941, p. 1073–1076) that "permissive intrusion, tectonically controlled" best fits the available evidence. Mayo believed that the forces governing intrusion were regional tectonic ones and that the plutons rose permissively rather than forcefully. Such a process seems to explain many of the broader relationships between granitic and metamorphic rocks, but not the intrusion of plutons into other plutons. It is difficult to explain the complex of Graveyard Peak, for example, in this way.

Zone melting as a mechanism of intrusion, newly proposed by Dickson (1958), might have been the dominant process. Dickson's theory, soundly based thermodynamically and chemically, is that a mass of magma, generated by any process but denied egress through fissures, would migrate upwards by melting its roof and crystallizing at its floor. This would result from water and other volatiles moving upward in response to hydrostatic-pressure gradients into the lowest-pressure portions of the chamber, thus lowering the crystallization temperature of the upper part of the magma and the melting temperature of the roof while raising the crystallization temperature and forcing consolidation of the lower part of the magma. Exothermic crystallization at the base would add heat, whereas endothermic reactions and heat loss at the top would subtract heat. Superimposed on this mechanism could be any combination of other processes, but both the room and superheat problems would vanish if zone melting were the dominant means of intrusion.

Many puzzling features of plutonic rocks can be explained in terms of Dickson's theory. The sharp and crosscutting yet passive-appearing contacts between plutons are easily accounted for, as are the scarcity of stoped blocks and the abundant evidence for major assimilation. The final composition of any pluton would be intermediate between its initial magma and the rocks intruded. Assuming the initial magma to be felsic, a magma intruding most types of metamorphic rocks would produce a more mafic final pluton. Younger plutons, intruding mostly granitic rocks, would be successively more felsic in final composition—the known relationship. None of the other mechanisms of intrusion yet proposed, alone or in combination, satisfy so many of the known conditions as Dickson's theory of zone melting.

AGE OF GRANITIC ROCKS: Numerous age determinations based on radioactive elements in granitic rocks of this part of the Sierra Nevada indicate a Middle Cretaceous age for the rocks. Seven zircon ages (Larsen lead-alpha method) for samples from near this area average 100 million years (Larsen *et al.*, 1954). Age determinations by the potassium-argon method of specimens from the Yosemite region northwest of this area range from 82 to 95 million years (Evernden, Curtis, and Lipson, 1957), and, if correct, show that successive adjacent plutons differ in age by something like a million years.

ORIGIN OF GRANITIC MAGMAS: The granitic rocks formed from hot, fluid intrusive magmas. The alaskite was so fluid that small mafic xenocrysts could sink to form gravity-graded layers, but this is unique, and in the other plutons small dark inclusions and mafic minerals show no sign of gravitational settling, although no floors were seen. The final motion in many plutons took place while they could deform without cataclasis yet had so high a viscosity that mineral and inclusion orientations could be preserved. The complex oscillatory-zoned cores and normally zoned margins common in plagioclase (Pl. 4, fig. 3) suggest a long period of crystallization in a mixing and convecting magma, followed by static crystallization of the sodic rims. Potassic feldspar and quartz characteristically crystallized late; the former replaced much plagioclase, but this replacement was apparently a normal feature of the crystallization and not a function of introduction from outside or below potassium or other material.

Fluids residual after partial crystallization of the granitic masses are the logical source of the aplites and pegmatites, and a filter-press process combined with the spontaneous expansion of the supercritical fluids seems an adequate mechanism for their migration.

Gravitative crystal settling is clearly shown by alaskite, but no evidence for it was noted in any other pluton. Even in alaskite there is no apparent compositional change with altitude except in the thin graded layers. The crude progression of composition with time within a group of plutons and the nature of variations in composition between diverse types within and between plutons are the only supports of a theory of differentiation. There are exceptions to the progression in composition, and other ways to explain it. The continuity of variations

expresses an empirical chemical relationship valid within practically any group of igneous rocks, whatever the origin of their diversity.

The gravitative crystallization differentiation postulated by Rogers (1958) for presumably correlative quartz monzonite in southeastern California is not substantiated, as the vertical zonation he claimed for steep-walled "Unit 1" pluton is not demonstrable. This pluton is 7 miles in diameter and crops out on a dissected pediment that has a local relief of 500 feet but an over-all slope of only 2° across the pluton. The small areas of good exposure are divided by Rogers into 3 subunits, each present in only one part of the pluton and going from bottom to top of all hills within that part, yet assumed to be due to vertical variations. The approximate altitude ranges of present exposures are: "lower" subunit, 3500–4300 feet; "transition" subunit (which nowhere appears between "upper" and "lower"), 3700–4100 feet; and "upper", 3900–4870 feet. In no hill does one type lie above the other, and the nearest exposures of the "upper" and "lower" subunits are almost 2 miles apart. The lithologic changes are easily accounted for as lateral variations (cf. Hamilton, 1956b)—indeed, Rogers' information shows the changes to be lateral rather than vertical.

The Sierra Nevada batholith, large though it is, is only a part of a great late Mesozoic batholith which is probably continuous from the tip of Baja California to northern Idaho. Within the batholith in southern California, and cut by related small plutons east of the batholith in California and elsewhere, are Precambrian plutonic rocks. The dominant Mesozoic granitic rock in such places is biotite quartz monzonite, which obviously formed from magmas derived from within or below the Precambrian plutonic rocks, beneath the geosynclinal prism of Paleozoic and early Mesozoic rocks.

In the area of this report, leucocratic biotite quartz monzonite is the dominant rock type, followed, in order, by granodiorite, granite and alaskite, quartz diorite, and diorite and gabbro. Although no chemical analyses are available, the mineralogy and relative abundances of the rocks indicate a frequency-distribution curve plotted against SiO_2 content would be strongly asymmetric, rising gently, then exponentially, from an origin near SiO_2 = 50 per cent, to a very high peak at about 73 or 74 per cent, then falling abruptly to end at about 80 per cent. Contamination by assimilation of metamorphic rocks can be demonstrated for each of the three plutons of granodiorite, and an origin of these rocks by contamination of felsic magmas is possible. The composition of the initial magmas was perhaps similar to that of the frequency peak. Although no field evidence within the area indicates that assimilation was important in the formation of quartz diorite, its origin is quite possibly similar, for the bulk of quartz diorites in the Sierra Nevada are in settings where large-scale assimilation can either be demonstrated or inferred (Hamilton, in press). Changes in composition between successive plutons might be due to both decreasing availability of contaminants and to complex sequences of fractional melting and crystallization.

The genetic relationship of metagabbro and metadiorite to the remaining rocks is uncertain. Metasomatic origins for such rocks have been postulated in other areas, but at least the northern mass of metagabbro in this area was probably intrusive, for it is a little-deformed and uniform mass completely different in composition from the surrounding low-grade and greatly deformed metasedimentary rocks.

Tertiary Volcanic Rocks

Basalts of Tertiary or early Pleistocene age cover about 1½ square miles of the south part of the area, where they lie as erosional remnants on a gently sloping ridge. A detailed field and petrographic study of these rocks is being made by P. A. Lydon. The basalts form a section of flows that in general dip west, roughly parallel to the slope of the ridge. The dominant rock type is aphanitic olivine-augite-labradorite basalt, in which the olivine and part of the augite form small phenocrysts in the dark aphanitic groundmass. With these dense lavas are interlayers and inclusions of scoriaceous olivine basalt, commonly reddish, and a little cinder-agglomerate. A small plug of basalt with vertical flow structures forms the mound marked by the 10,179-foot altitude in the western area of basalt, and there are many squeeze-ups in the lavas elsewhere. The north end of the northeast area of basalt is marked by 400-foot cliffs of basalt jointed in continuous near-vertical columns. The lower contact of these rocks is irregular but was not visited, and it is uncertain whether the cliffs expose a volcanic neck or a single extremely thick flow that was erupted over rough topography.

A vertical dike of vesicular basalt, 2 feet thick and striking north-northwest, crosses Minnow Creek about two-thirds of a mile

north of Marilyn Lake, near the northwest corner of the area.

Because the basalt flows are positioned only on a broad ridge and because of the deformation represented by their westerly dip, they are considered to be of Tertiary age; certainly they are older than the incision of the present valleys. The basalts might, however, be early Pleistocene.

Surficial Geology

Repeated glaciation during the Quaternary profoundly affected the topography. All the area except the higher summits was covered by ice during at least one of the Pleistocene glacial stages, and the peaks that protruded above the ice were left as horns, sapped from all sides by cirques.

The glacial deposits shown on Plate 1 are for the most part those that have marked topographic expressions. Most of them are of probable Wisconsin age. The glacial geology of much of the Mono Creek drainage was studied by Birman (1954); the Rock Creek and Hilton Creek drainages were studied by Webb (1956) and by P. C. Bateman. No complete reports have been published, but the map accompanying Birman's preliminary report (1954, Pl. 1) includes much of the southwest part of this area. The oldest morainal debris (Sherwin?, which is pre-Wisconsin) has little or no topographic form and consists of till containing deeply weathered erratics. The several sets of Wisconsin moraines have, in general, well-preserved form; Birman distinguished three sets, Tahoe, Tioga, and intermediate. Webb distinguished two. The early moraines go farther down the canyons and higher up the valley sides than the late moraines, but all extend beyond the limits of this area.

The extensive Wisconsin moraines of the southwest part of the area include well-shaped lateral and medial moraines and areas of irregular ground moraine. Before construction of the Edison Lake dam, well-preserved upper Wisconsin recessional moraines were exposed in the southwest half of the reservoir area. Upstream from these and underlying most of the rest of the reservoir site was a broad valley floor of lenticular and cross-bedded outwash sands and gravels. The moraines were largely destroyed during excavation for the dam, and great quantities of sediment were scooped also from the outwash deposits. The glacial deposits in the extreme northeast corner of the area belong to the Tioga (late Wisconsin) stage, according to P. C. Bateman (personal communication).

The glacial deposits lie chiefly in the lower parts of the area. The uplands are of bare fresh rock, sprinkled sparsely by erratics but covered by till only locally. (Small areas of surficial deposits are not shown on Plate 1.) The late Wisconsin ice, flowing over a bedrock surface scoured by ice not long before, left an exceedingly clean terrain, and apparently the late ice had little rock debris available to it. Cirques, barren basins, serrated ridges, and glacial troughs are the main elements of the landscape.

Tiny and dwindling glaciers are present on the sheltered north slopes of four high peaks in the southeast part of the area and reach a minimum altitude of 12,000 feet. In Recent time, many of the cirques were occupied by small glaciers, some of them a mile or more long. The small moraines left by these Recent glaciers are unweathered, and some are unstable underfoot. Birman (1954) distinguished three sets of such moraines.

Post-Wisconsin talus debris mantles many glacially oversteepened slopes. Below cliffs, the distinction between talus and bedrock is obvious, but on many gentler slopes talus grades upward into discontinuous rock rubble. The biggest talus slopes are those along Mono Creek, locally bare and obviously active but generally covered by forest or scrub.

Above timber line, where frost occurs nightly during most of the summer, talus formation and other alpine geomorphic processes are currently active, and frost-riven blocks cover many ridges. Several rock glaciers, two of them more than half a mile long, fill cirques north of Mt. Crocker and Mt. Stanford.

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STRUCTURAL SIGNIFICANCE OF GRAVITY SURVEYS IN THE VIRGINIA CITY—MOUNT ROSE AREA, NEVADA AND CALIFORNIA

BY GEORGE A. THOMPSON AND CLARENCE H. SANDBERG

ABSTRACT

Gravity surveys were made in the Mount Rose and Virginia City quadrangles to investigate the concealed Cenozoic structure of the alluviated basins and to explore for significant gravity differences between basins and ranges. The gravity data have been computed as complete Bouguer anomalies, which incorporate corrections for all the accurately assessible factors affecting gravity values, including topographic irregularities but not isostatic compensation.

Gravity variations in the basins are closely related to changes in thickness of Cenozoic sedimentary rocks and alluvium in the basins. The gravity data indicate that light Cenozoic sedimentary rocks in Washoe Valley are at least 1800 feet thick; that the sediments in the northern part of Steamboat Valley are only about 600 feet thick and thin gradually eastward but thin abruptly to the west near Steamboat Hills; that east of Reno, between Sparks and the Virginia Range, sediments are roughly 2800 feet thick and thin abruptly about half a mile west of the range; and that in the basin west of Reno sediments are locally 1800 feet or more thick. The variations in thickness reflect important structural deformation in the basins. The late Cenozoic structures that separate the basins from the mountains are a complex of contemporaneous faults and folds; folding is predominant in some areas and normal faulting in others.

From the eastern boundary of the area surveyed, the gravity values decrease westward about 1 mgal per mile, a total of 30 mgals, to a minimum at the east side of the Carson Range of the Sierra Nevada and then increase toward the main Sierra block. There is little or no gravity expression of the Virginia Range or the Carson Range relative to sediment-free parts of the basins adjacent to the ranges.

Gravity measurements of the Coast and Geodetic Survey combined with the present measurements demonstrate that the Bouguer anomaly from San Francisco eastward shows a general inverse relationship to elevation but is a minimum along the east side of the Carson Range rather than at the main summit of the Sierra. This gravity profile may be the result of a thickening of the earth's crust eastward beneath the higher part of the continent, combined with a smaller effect from greater-than-average amounts of granitic rock in the crust of the Sierran region.

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Introduction

The Virginia City-Mount Rose area lies across the boundary between the Sierra Nevada and the Basin Range Province. The main mountain block of the Sierra lies just west of the area; the Carson Range of the Sierra (Pl. 1) is an offshoot extending northward from the main block and separated from it at this latitude by the Tahoe-Truckee trough; the Virginia Range is one of the Basin Ranges.

The geology of the Mount Rose and Virginia City quadrangles, which are the two 15-minute quadrangles covering the major part of the area (Pl. 1), is generalized from maps by Thompson (1956; in press). The outstanding contribution of Calkins (1944) and Calkins and Thayer (1945) to the geology of the Comstock Lode is incorporated in these quadrangle maps. The geology of the part of the area lying in California is taken from Lindgren (1897) and of the part south of Washoe Lake from Reid (1911). Gianella (1936), working in the area south of Virginia City, made the first modern study of the volcanic rocks. White and associates (Brannock and others, 1948; other reports in manuscript) have made a thorough study of the Steamboat area. King (1878), Becker (1882), Louderback (1904; 1906), and many others made earlier contributions to the geology of the area.

Gravity measurements were undertaken to investigate the concealed structure in the alluviated basins, to explore for significant gravity differences between basins and ranges, and to study the eastward extent of the theoretical compensating root of the Sierra Nevada.

The only previous gravity measurements that have been published are the pendulum gravity determinations made by the Coast and Geodetic Survey along U. S. Highway 40 over the Sierra Nevada (Duerksen, 1949; Johnston, 1940). In their well-known earthquake studies, Byerly (1938; 1939) and Gutenberg (1943) agree in interpreting the delay in earthquake waves passing beneath the mountains as caused by a root of crustal rocks extending into the mantle of the earth. This interpretation was challenged by Mintrop (1953), but his conclusions are based partly on data from stations far from the mountains.

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Profs. Adolf Knopf and Konrad B. Krauskopf of Stanford University have helped this work with stimulating discussions of the geologic problems. The writers are especially indebted to Donald E. White of the U. S. Geological Survey for continuing enthusiastic support, for data on the geology of the Steamboat Springs area, and for much help in field work and interpretation. P. E. Byerly, Jr., assisted with the measurements and reduction of the data in the early stages of the gravity investigation. In the work reported here the gravity measurements and computations are mainly the responsibility of Sandberg; the geologic interpretations are mainly the responsibility of Thompson.

Gravity Measurements

General Statement

About 200 gravity stations were established in 1952 in an area of 600 square miles, an average of 1 station for every 3 square miles. With the exception of traverse B-B′ (Pl. 1), all gravity measurements were made with a Frost gravimeter having a sensitivity of 0.0734 mgals per scale division and a working range of 1500 scale divisions. Traverse B-B′ was run by a Stanford University geophysics class using Stanford's gravity meter, similar in type to the Frost but with a sensitivity of 0.1263 mgals per scale division. Since a gravimeter measures only differences of gravity, all stations were referred to the pendulum gravity stations of the Coast and Geodetic Survey at Truckee and Mystic (Duerksen, 1949). The final values are expressed as Bouguer anomalies.

Bouguer Gravity Anomalies

The Bouguer anomaly as used in this paper is defined (in milligals) as

$$g_{T_r} + g_d + 0.05998\,h + T - g_\theta$$

g_{T_r}	g_d	$0.05998\,h$	T	g_θ
Gravity at Truckee (USC & GS)	Observed difference from Truckee	Elevation and simple Bouguer corrections; h is elevation	Terrain correction	Theoretical gravity at sea level

The Bouguer anomalies thus incorporate corrections for all the accurately assessible factors affecting gravity values but not for isostatic compensation, which may be regarded as part of the unknown mass distribution to be determined. The advantage of the Bouguer anomaly over isostatic anomalies is that the latter incorporate assumptions of particular mass distributions (Pratt vs. Airy, with various depths of compensation; local vs. regional). Though isostatic anomalies are extremely useful in broad regional studies, these assumptions make their use restrictive in a detailed analysis of a small area. If the area is one of large faults cutting the earth's crust into blocks, compensation is possible with the crust depressed deeper ("roots") beneath graben blocks than beneath horst mountain blocks. This means that the difference in isostatic anomaly between a mountain and a basin may vary with the type of isostatic anomaly computed, and if crustal-shear isostatic anomalies are included (Vening Meinesz, 1948) the difference may vary not only in magnitude but even in sign. The free-air anomaly is also unsuitable; it is equivalent to an isostatic anomaly computed on the basis of local compensation at negligible depth.

Bouguer anomalies are plotted as an isogal map (Pl.1) with a contour interval of 5 mgals.[1] All values are negative, as is appropriate for a high continental area that is compensated; they range from −167 to −208 mgals.

The precision of the field measurements is believed to be within 0.2 mgal for most of the stations and within 0.5 mgal for all stations. Pendulum base stations to which the survey is necessarily tied have an uncertainty relative to each other of more than 1 mgal (Duerksen, 1949, p. 2–4; Woollard, 1950). Elevations of most of the stations are known to the nearest foot, permitting elevation corrections within 0.1 mgal. A specific gravity of 2.67 was used in correcting for the attraction of material between a station and sea level, following the practice of the Coast and Geodetic Survey. All differences between 2.67 and the actual specific gravities down to sea level thus contribute to the Bouguer anomaly. Where there is large topographic relief within an area of rocks uniformly lighter or heavier than 2.67, the gravity anomaly will vary depending upon the elevation. For example, the gravity anomaly on a hill in an area of uniformly lighter rocks will be negative relative to an adjacent valley, correctly reflecting the greater thickness of light rocks between the top of the hill and sea level.

The topographic correction out to a radius of 13.6 miles (the M zone of the Hammer Tables) was about 2 mgals at most of the stations and a maximum of 10 mgals at one station. A specific gravity of 2.67 was used in making the correction, because this specific gravity is reasonable for the mountain masses that make the main contribution to the correction. The topographic corrections involve the largest uncertainty of any of the corrections, but they are reproducible with a precision of better than 0.5 mgal.

The theoretical sea-level value of gravity at each station is that determined from the International Gravity Formula of 1930.

Specific Gravity of Rocks

The specific gravities of representative rocks were determined in the laboratory in connection with petrographic studies (Table 1). The determinations for fresh granitic and metamorphic rocks are consistent and reliable. They show that the metamorphic rocks are denser than the granitic rocks. An average specific gravity of the granitic and metamorphic rocks is close to 2.67, the value used by the Coast and Geodetic Survey for gravity computations.

The specific gravities of the fresh volcanic and sedimentary rocks vary widely, depending partly on mineralogic composition and to a larger extent on porosity. Laboratory measurements give only a very rough indication of the range for these rocks and also for the altered rocks. Field determinations of specific gravities of these rocks from the gravitational effects of topography are also of limited significance because of wide variations in specific gravity over short distances horizontally and vertically and the difficulty of finding suitable topographic features in the right places. The average specific-gravity contrast between alluvial sediments and bedrock in Steamboat Valley was estimated from the evidence in drill holes to be approximately −0.5.

Concealed Structure of the Basins

Introduction

The structure of the alluviated basins is an unusually favorable problem for gravity measurements because of the large density contrast between bedrock and basin sediments and

[1] The gravity measurements and corrections are on file at the Geophysics Branch, U. S. Geological Survey, Menlo Park, California.

because of the moderate depths to bedrock. The borders of the basins in many places are pediments with a thin veneer of alluvium, which conceals the structural breaks between mountains and basins. Gravity determinations give a measure of the thickness of basin sediments and hence indicate the location and character of the structural breaks. Similarly the thickening and thinning of sediments within the basins delineate faulted, tilted, and warped blocks.

Any interpretation of gravity data without other information must be ambiguous, but no geologic interpretation can be correct that does not account for the gravity measurements. If the geologic possibilities are limited to several solutions, gravity data may eliminate some and show which is correct. For example, if we wish to find out whether a concealed basin floor is steeply faulted down or gently tilted down, gravity data can generally give a clear answer. It is with questions of this sort that this paper is mainly concerned.

TABLE 1.—SPECIFIC GRAVITY OF ROCKS

The specific gravity of water-saturated rocks is computed as $W(P - L)/P + L$, where P is the powder specific gravity, L is the lump specific gravity, and W is the specific gravity of ground water, 1.0.

Rock	Number of samples measured	Specific gravity of water-saturated rock	
		Range	Average
Truckee formation	1	1.74–	
Volcanic rocks of Cenozoic age (quantitatively most important groups italicized)			
Lousetown formation	11	2.49–2.80	2.66
Kate Peak formation, flows	7	2.53–2.69	2.61
Kate Peak formation, tuff-breccia	3	1.84–2.18	2.04
Kate Peak formation, intrusions	6	2.41–2.55	2.50
Kate Peak formation, vitrophyre	3	2.13–2.42	2.31
Alta formation	8	2.54–2.70	2.63
Alta formation, Sutro member	1	2.50–	
Hartford Hill rhyolite tuff	4	1.95–2.56	2.25
Davidson granodiorite (Tertiary)	2	2.64–2.67	2.65
Pre-Tertiary granitic rocks	5	2.62–2.66	2.64
Metamorphic rocks	3	2.69–2.75	2.71
Altered granitic rocks	6	2.46–2.62	2.55
Altered volcanic rocks	21	2.07–2.72	2.50

Washoe Valley

The largest negative anomaly, −208 mgals, is in the valley occupied by Washoe Lake (Pl. 1). The valley is partly bounded by faults on both the eastern and western sides. Late Cenozoic deepening of the basin is indicated by the presence of the lake, by uplifted basin sediments at the north end, and by a peculiar entrenched outlet. The gravity anomaly suggests that the basin is asymmetrical and that its deepest part is closer to the west side. The thickness of basin sediments can be estimated from the local anomaly, which is −10 to −15 mgals relative to bedrock areas surrounding the basin, by the formula $T = A/0.013S$ mgals[2], where T is the thickness in feet, A is the local anomaly in mgals, and S is the difference in specific gravity between the sediments and bedrock. If the contrast in specific gravity is assumed as −0.5, and the local anomaly is −12 mgals, the thickness is about 1800 feet. The calculated thickness is inversely proportional to the assumed contrast in specific gravity, so that the thickness may be much greater but not much less than 1800 feet. A small error is introduced by treating the mass of sediments as an infinite slab, and this error tends to minimize the calculated

[2] The attraction of an infinite slab is 0.013 mgal per foot of thickness per unit specific gravity.

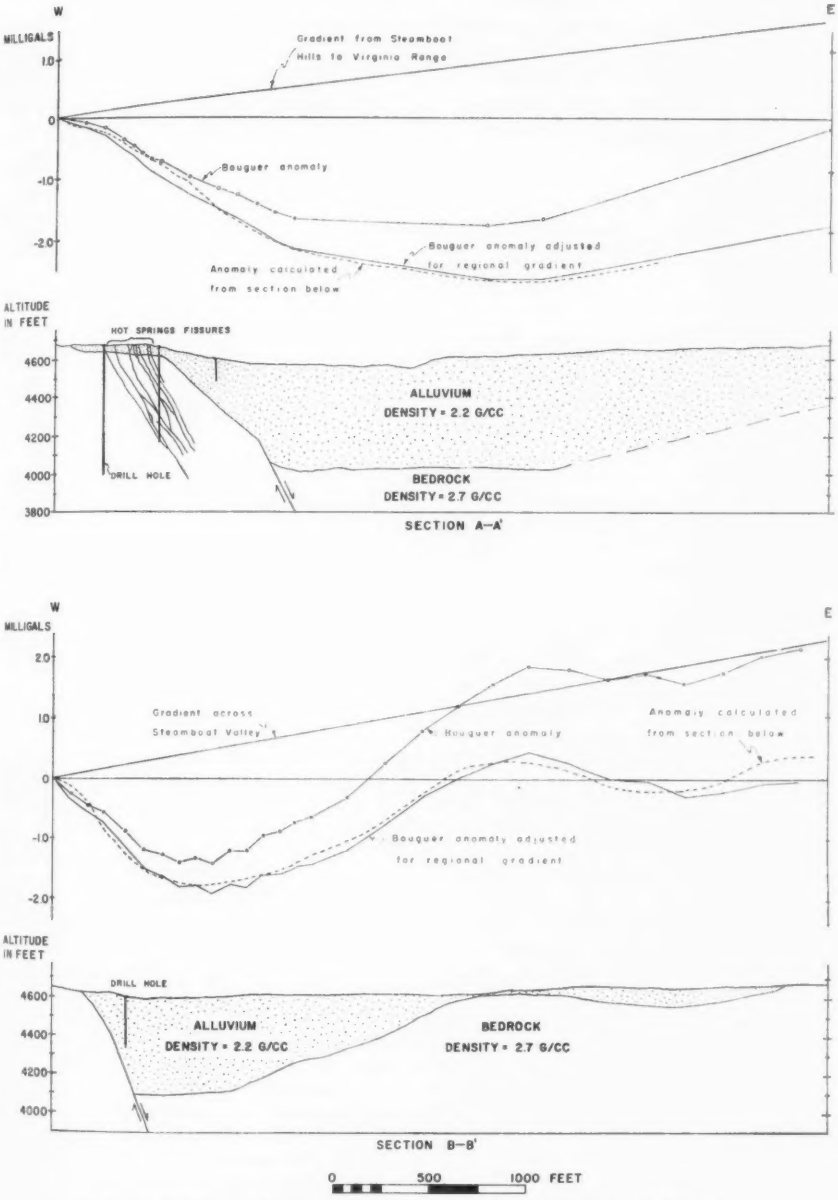

FIGURE 1.—GRAVITY PROFILES AND GEOLOGIC SECTIONS, STEAMBOAT VALLEY
Geology partly from D. E. White (Unpublished data)

thickness of sediments. More refined calculations are not warranted by the available data, but the rough calculations indicate a thickness of 1800 feet or more of sediments in Washoe Valley.

Steamboat Valley and Southeastern Truckee Meadows

In the east end of Steamboat Hills an extensive system of hot springs issues from fissures

(Brannock and others, 1948; White, 1955). Two lines of gravity stations were established eastward from the springs.

The Bouguer anomaly along line A-A′ (Pl. 1) is shown in Figure 1 with a geologic section determined by assuming various sections, within the limits imposed by geologic control, and adjusting to fit the gravity. Geologic control is provided by five drill holes, three of which are on the line of traverse. A gravity profile computed from the geologic cross section is shown by short dashed lines for comparison with the observed anomaly. Drill-hole information was not sufficient to establish the thickness of alluvium under the valley floor, or to determine the amount of displacement of the faulted bedrock surface. By use of gravity and drill-hole data we can determine that the maximum thickness of alluvium is about 600 feet, that there is probably no pronounced displacement of the bedrock surface beneath the prominent fissures from which the hot springs issue, and that there is no more than a small vertical displacement of the bedrock surface at the fault. Large movements on this fault, indicated by the geologic relations at the southern end of Steamboat Valley (Thompson, 1956), must have taken place before the alluvium was deposited.

The second gravity profile, B-B′ (Fig. 1), about a mile south of A-A′, crosses the same fault. The gravity gradient at the west end is steeper than the gradient on profile A-A and indicates less erosion of the fault scarp before deposition of alluvium. From a maximum

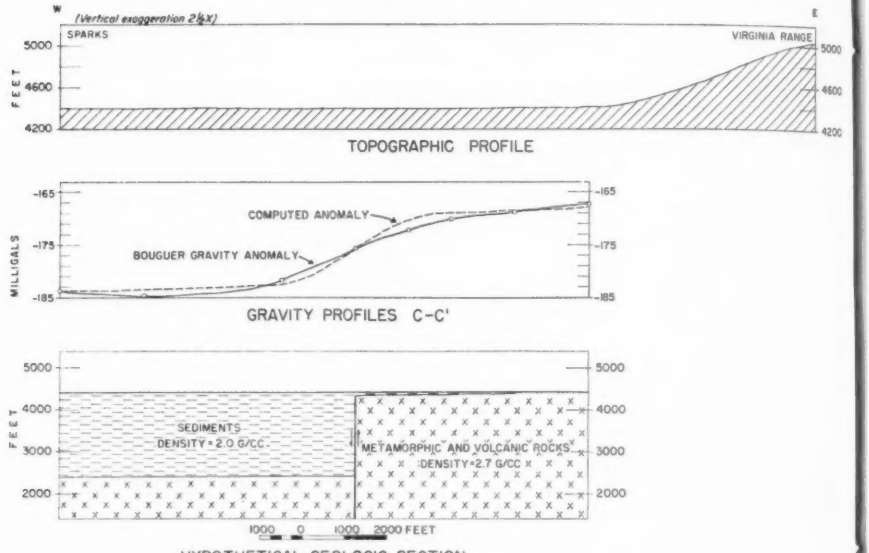

FIGURE 2.—PROFILES AND SECTIONS BETWEEN SPARKS AND THE VIRGINIA RANGE

thickness of about 500 feet near the fault, the alluvium seems to thin gradually eastward, which suggests that the valley block has been tilted westward.

Truckee Meadows

North of Steamboat Hills is Truckee Meadows, a broad alluviated basin (Pl. 1). Hills of bedrock surrounded by sediments and numerous small faults in the basin sediments themselves testify to the structural complexity of this valley. Borders of the basin are generally irregular, and the structures at the base of the mountains are concealed beneath younger sediments.

Northeast of Steamboat Hills the gravity anomaly is a minimum for the basin, −190 mgals. This anomaly is about 10 mgals more negative than that on near-by areas of bedrock. The indicated thickness of sediments, by the same line of reasoning as that used for Washoe Valley, is roughly 1000 feet, or possibly considerably more if the specific-gravity contrast is less than 0.5. A water well in the south part of the anomalous area is reported by D. E. White (Oral communication) to be 680 feet deep and entirely in sediments.

About 3 miles southeast of Reno is another gravity minimum in which gravity values are about 10 mgals more negative than those on the nearest bedrock. This anomaly indicates at least 1000 feet of sediments. Between this gravities. As the specific gravity of the sediments varies with depth in an unknown manner, a refinement of the computation is not worthwhile.

Limits to the indicated thickness of sedi-

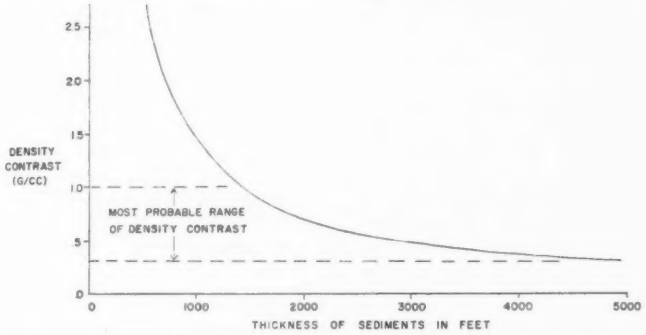

FIGURE 3.—THICKNESS OF SEDIMENTS PRODUCING AN 18-MGAL GRAVITY ANOMALY, AS A FUNCTION OF DENSITY CONTRAST

gravity minimum and that northeast of Steamboat Hills, bedrock in Huffaker Hills rises 500 feet above the valley. This elevation, taken together with the depth of valley fill computed from the gravity minima on either side of the hills, indicates a total relief of at least 1500 feet on bedrock. Probably the bedrock relief represents structural relief along faults or folds trending northeastward, following the trend of the bedrock hills and that of known faults in the hills.

An unusually steep gravity gradient characterizes the area east of Reno between Sparks and the Virginia Range (Pl. 1). In Figure 2, topography, exaggerated 2½ times, is shown in the top profile to locate the gravity anomaly relative to the Virginia Range, and directly below is plotted the gravity profile (C-C'). Gravity values change only gradually for more than half a mile west of the Virginia Range, then decrease abruptly, and finally flatten out farther west. The change in gravity amounts to about 18 mgals, and on the assumption of a specific-gravity contrast of 0.5, would require a sheet of basin sediments about 2800 feet thick to account for the anomaly.

If the difference in specific gravities were 0.7 and the contact were vertical, as shown in the lower cross section of Figure 2, the thickness would be approximately 2000 feet, and the gravity anomaly would be that shown by the short dashed line. Close agreement with the observed gravity anomaly can be obtained by modifying the dip of the postulated fault contact and altering the assumed specific ments are fixed by the possible range in density contrast. If the sedimentary basin were filled with water instead of sediments (specific-gravity contrast 1.7), the depth would still be more than 800 feet (Fig. 3). Although a density contrast of 1.7 is absurd, the contrast might be as great as 1.0 because thick sections of very light diatomaceous rocks crop out west of Reno; if the contrast is 1.0, the thickness of sediment would be approximately 1400 feet. It is unlikely that the contrast is less than 0.3, for which the thickness of sediment would be about 4600 feet (Fig. 3). The most probable value is a contrast of about 0.5, from which a thickness of 2800 feet is computed.

Regardless of the exact thickness, the gravity data indicate a major structural break, not at the base of the Virginia Range, but about a mile to the west. Between this break and the range is an erosion surface cut on bedrock and thinly alluviated. The structure is interpreted as a fault rather than a monoclinal fold because the basin sediments, where they have been exposed by excavations in Truckee Meadows, are nearly flat, and the gravity gradient is the steepest that was found along any of the basin margins. Although gravity stations are too few in some areas (as along the west side of Washoe Valley) to constitute a conclusive test, the gradients are generally smaller and probably mean that the basins usually have gentler sloping margins than east of Sparks. Gentler margins occur where the basins are bounded by tilted or folded bedrock, where fault scarps

have been extensively eroded before being buried by basin sediments, or where displacement is distributed among a number of separate faults.

Basin West of Reno

In the valley of the Truckee River west of Reno, the Truckee formation is many hundreds of feet thick and consists of shale, diatomite, and andesitic sandstone. These rocks lie with only slight angular discordance on the volcanic rocks of the Carson Range and dip northeastward and northward from the range. Fairly gentle folding, only slightly modified by faults, thus defines the southern margin of this large body of sedimentary rocks.

The gravity anomaly (Pl. 1) is −192 mgals close to Reno on the southwest side, about 12 mgals lower than the nearest values on bedrock. If a specific-gravity contrast of 0.5 is assumed, the anomaly indicates a thickness of roughly 1800 feet of sediments. An exploratory drill hole southwest of Reno (Anderson, 1910, p. 487) penetrated 1890 feet of sediments.

The western side of the gravity minimum, 3 miles west of Reno, shows a steep gradient and clearly indicates westward thinning of the sedimentary rocks. From here westward to the California State line, the gravity data suggest that the sedimentary rocks are no more than a few hundred feet thick.

Other Basins

Near the town of Truckee (Pl. 1) is another gravity low of about −190 mgals. Gravity data are sparse but suggest a correlation of the local gravity low with sedimentary rocks. The Truckee formation in this area is partly covered by younger lava flows, so that a close correlation with the outcrop area of the Truckee would not be expected.

Along the north shore of Lake Tahoe, valleys are deeply alluviated and, as would be expected, show gravity lows. Most of the strip of alluvium crossing the Carson Range northeast of Lake Tahoe is glacial outwash and moraine and is so thin that it hardly affects the gravity measurements.

In the basin occupied by Carson City, the gravity stations are too few to permit separation of the effects of basin sediments from regional effects of the basement rocks, but the data suggest that sediments in this basin are only moderately thick.

The valley east of the Virginia Range is bordered on the west side by major faults, partly concealed at the base of the range. Gravity values of about −185 mgals near the southwest side of the valley are roughly 10 mgals lower than the values on the nearest bedrock and suggest that the sedimentary fill is about 1500 feet thick.

In the interior of the Virginia Range 6 miles northeast of Virginia City, the Truckee formation covers an area of several square miles known as Chalk Hills. Two gravity stations, which are located near one edge of the area of sedimentary rocks, show values about 4 mgals lower than those on near-by bedrock. The indicated thickness of the Truckee formation is thus several hundred feet at the location of the stations.

Borders of the Basins

The late Cenozoic structures that separate the basins from the mountains are a complex of contemporaneous faults and folds; in some areas folding is predominant, in others, normal faulting (Thompson, 1952). The fault pattern is not shown on Plate 1 because a generalization from maps of larger scale (Thompson, 1956; in press) would be misleading. The main facts that correlate with the gravity data, however, can be readily indicated.

The lowland between the Carson Range and the Virginia Range, extending through Reno on the north and Carson City on the south, is conspicuously composed of a chain of smaller individual basins, each partly or wholly divided from the others by saddles that connect the two ranges. From north to south the principal individual basins are (1) the northern part of Truckee Meadows, bounded on the south by Huffaker Hills, (2) the southern part of Truckee Meadows, lying between Huffaker Hills and Steamboat Hills, (3) Washoe Valley, and (4) the basin in which Carson City lies. The northern part of Truckee Meadows is bounded on the east by a concealed normal fault, which is indicated by the gravity measurements (Fig. 2). To the south this fault dies out toward Huffaker Hills. The Truckee formation south and southwest of Reno strikes generally parallel to its contact with the underlying volcanic rocks and dips to the northeast; a fold thus limits this side of the basin. The southern part of Truckee Meadows is bounded by folds on both the east and west sides. The volcanic rocks in the Virginia Range northeast of Steamboat Hills dip westward 30°, and the Truckee formation northwest of Steamboat Hills dips

eastward 30°. The folded margins of this basin are consistent with the smaller gravity gradients. The Steamboat Hills block is a faulted fold or strongly warped fault block. Washoe Valley is bordered on the west by a great fault and on the east by a fold and smaller faults. The basin of Carson City may be bordered on the west by a fold or by distributed faults, as indicated by the small gravity gradient on the east side of the Carson Range southwest of Carson City.

The structural character of the east front of the Carson Range thus varies along the chain of basins: a great fold southwest and south of Reno, a fault at Washoe Valley, a probable fold near Carson City, and beyond this area to the south a large fault again. The western side of the Virginia Range also varies in structural character along the chain of basins.

Sierra Nevada and Virginia Range

Correlation of Gravity with Rocks in the Ranges

Correlation of gravity with granitic and metamorphic rocks in the ranges is notable; gravity is generally lower on the granitic rocks. For example, on the west side of the Virginia Range east and northeast of Washoe Valley (Pl. 1), the gravity contours follow the trend of an intrusive contact and define a fairly steep gravity gradient sloping downward toward the granite area at the rate of about 5 mgals per mile. The metamorphic rocks there contain much debris of intermediate and basaltic volcanic rocks and are denser than the granite. Metavolcanic rocks crop out in the Virginia Range near the north and south edges of the map (Pl. 1). In both places the gravity values are unusually high, −170 and −175 mgals respectively. A gravity high (−170 mgals) on granitic rocks at the eastern edge of the map is probably explained by the general increase (regional gradient) of gravity to the east. Finally, gravity is unusually low throughout the large area of granitic rocks in the Carson Range.

Specific-gravity measurements on the granitic and metamorphic rocks (Table 1), though too few for broad generalizations, show consistently smaller values for the granitic rocks and are thus in good accord with the observations of gravity.

The wide range of specific gravities of the volcanic rocks (Table 1) has less effect on gravity than might be expected, for at any one place the volcanic rocks are likely to be a mixed pile of light and heavy rocks. The gravity low 2 miles north of Virginia City possibly represents a local abundance of less dense pyroclastic rocks.

Altered rocks at Virginia City and on the west side of the Virginia Range northwest of Virginia City produce no noticeable local anomaly (Pl. 1), although the specific gravities of the exposed altered rocks are low (Table 1). This lack of local anomaly is evidence of the superficial character of the intense acid leaching (Thompson, 1956); at depth, pores are filled, and the altered rock is slightly denser than fresh rock. In Table 1, the highest specific gravities among altered volcanic rocks are those of samples from deep drill holes and mine workings.

Ranges Compared to Basins

A comparison of Bouguer anomalies in the ranges and in sediment-free parts of the basins is of considerable geologic interest. Bedrock connections between the ranges and bedrock exposures along the axes of the basins (Pl. 1) make this area well suited for significant comparisons, because the effects of low-density sedimentary fill in the basins can be eliminated by considering only the gravity stations on bedrock.

If the Carson and Virginia ranges are partly or wholly compensated isostatically they might show local negative Bouguer anomalies compared to the adjacent basins, reflecting low-density compensating rocks beneath the mountains. If the ranges were formed by simple uplift with no regard for the requirements of isostasy, they might show local positive Bouguer anomalies, owing to uplift of heavier, deep-seated rocks below the ranges. Evidence for a lack of local isostatic compensation of individual basins and ranges was long ago advanced by Gilbert (1890, p. 357–360, 365–392) and by Gilluly (1928, p. 1123–1130). Gilbert studied the broad warping of the basin of Lake Bonneville in response to filling and partial emptying of the lake and noted a lack of resulting fault movements along the base of the ranges. However, nonvertical normal faults cannot move without horizontal extension of the strata, and the required horizontal movement is surely independent of loading and unloading by a lake. Consequently, the lack of immediate response by faulting may not be a conclusive test of approach to compensation.

Vening Meinesz (1948, p. 20; 1950) studied isostatic compensation of normal faulted areas.

His results, like those of LeConte (1889), show that horsts and grabens may be in isostatic balance simply because of the geometry of the blocks and the increasing density of rocks with depth in the earth. The narrow base of a graben block requires it to sink deeper than a horst block to reach buoyant equilibrium. Grabens would be expected to show local negative Bouguer anomalies, and horsts local positive anomalies. Vening Meinesz thus explained the negative anomalies associated with the African rift valleys and eliminated any compelling reason for accepting Bullard's (1936) interpretation of the gravity data as indicating reverse faulting. A horst mountain in isostatic equilibrium may show the opposite Bouguer gravity anomaly from the usual isostatically balanced mountain.

Unfortunately from the standpoint of testing these hypotheses, the Carson and Virginia ranges show no significant gravity contrasts, either positive or negative, with adjacent basins (Fig. 4). After allowance is made for minor undulations of the gravity profile, which are too steep to be of deep-seated origin, and for known basin sediments, the ranges and near-by basins are gravimetrically almost the same. The results are the same as if the mountains were excess loads piled up on an unyielding surface. Though the Virginia Range owes its relief partly to a thick pile of volcanic rocks (Thompson, 1956), much of the Carson Range is free of volcanic rocks. Possibly the denser rocks of the deep crust and subcrust lie so far down in comparison with the narrow width of the mountains and basins that the gravity contrast to be expected amounts to only a few milligals. Such a small difference cannot be detected against the background of minor undulations related to rocks near the surface.

Regional Gradient and Sierran Root

The largest negative Bouguer anomalies in the area surveyed lie along an axis trending approximately north-south on the eastern side of the Carson Range. This statement also holds true in considering the stations on bedrock only, thus eliminating the effects of the light alluvial fill. Figure 4 shows the topography and Bouguer anomaly in a section drawn approximately east-west. The local gravity lows near each end of the gravity profile correspond to areas of alluvium and light sedimentary rocks. The sharp gravity high near the crest of the Virginia Range seems best interpreted as a response to dense metamorphic rocks, concealed at this point but exposed farther south in the Virginia Range. The steep gravity gradient at the center of the profile is correlated with the intrusive contact of granitic with metamorphic rocks. If all these local irregularities are eliminated, the smoothed gravity profile is as shown by the dashed line in Figure 4. The regional minimum lies along the east side of the Carson Range.

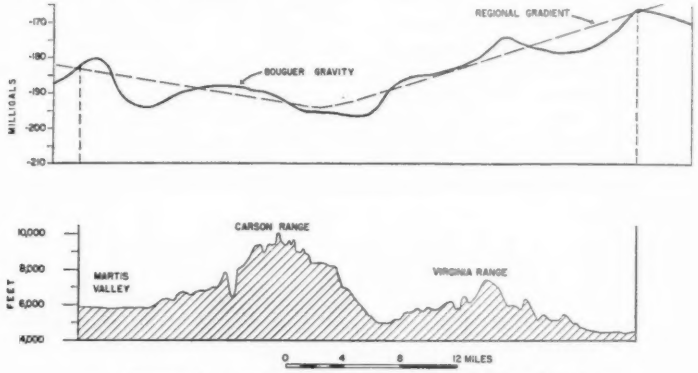

FIGURE 4.—GRAVITY AND TOPOGRAPHIC PROFILE D-D'

The gravity picture can be extended westward across the Sierra Nevada by using measurements made by the Coast and Geodetic Survey (Duerksen, 1949; Johnston, 1940). Figure 5 shows the topography on a line between San Francisco and Reno and the gravity anomalies projected to the same line. The Bouguer anomaly shows an inverse relation to elevation, which indicates a general tendency toward isostatic compensation. From the Sacramento Valley eastward, the Bouguer

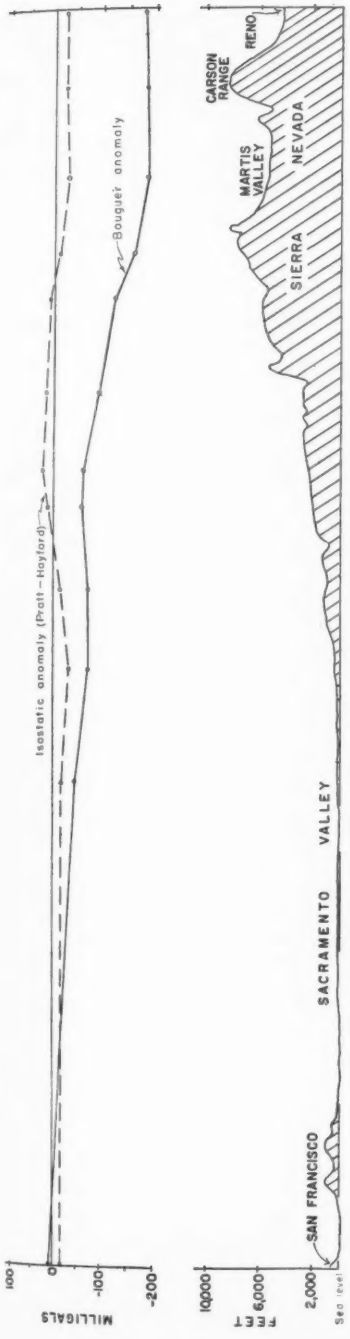

FIGURE 5.—GRAVITY AND TOPOGRAPHIC PROFILE, SAN FRANCISCO, CALIFORNIA, TO RENO, NEVADA

Topographic profile is along a straight line from San Francisco to Reno; gravity profile is along U. S. Highway 40 projected onto topographic profile. Gravity data from Coast and Geodetic Survey (Duerksen, 1949). Scale: 1 inch = about 25 miles.

anomaly becomes more negative toward the main summit of the Sierra Nevada. Farther east, the negative trend continues toward Martis Valley, in striking opposition to the usual change with elevation. From Martis Valley across the Carson Range to Reno the Bouguer anomaly remains near −180 mgals. The profile in Figure 4 shows the regional gradient continuing downward to the east from Martis Valley and reaching a value of −195 mgals along the east side of the Carson Range, where the gradient finally reverses. The lack of correspondence between anomaly and elevation east of the main summit is also demonstrated by the negative isostatic anomaly of about −25 mgals extending from Martis Valley to Reno (Fig. 5).

The simplest interpretation of the anomalies is that a large volume of low-density material with its edge in the western foothills of the Sierra Nevada increases to a maximum amount at the eastern side of the Carson Range and then decreases farther east. The low-density material may be in the form of (1) lighter-than-average rocks within the crust of the earth, (2) lighter-than-average material below the M discontinuity, within the mantle of the earth, or (3) a root of crustal material extending downward into the mantle of the earth (deeper M discontinuity). Combinations of these three solutions, or more complicated distributions, are of course possible.

The first solution may be visualized as higher-than-average amounts of granitic rocks in the crust. Worzel and Shurbet (1955, p. 91–92) computed an average specific gravity of the crust of 2.84 from gravity measurements and seismically measured thicknesses of continental and oceanic crust. The average specific gravity of granitic rocks is close to 2.67. A Bouguer anomaly of −200 mgals would therefore require a thickness of granitic rocks approximately equal to $200/42(2.84 - 2.67)$[3] or 28 km. Nearly the entire thickness of "average" crust would have to be replaced by granite to explain the observed anomalies. The smaller negative anomalies (−170 mgals) east of the Virginia Range would require about 24 km of granitic rocks in place of average crust.

The second solution, lighter-than-average material in the mantle, postulates conditions so far removed from geologic observations, so independent of gravity observations, and so

[3] The attraction of an infinite slab is 42 mgals per km of thickness per unit specific gravity and is a fair approximation for the attraction of a finite slab that is shallow in comparison to its horizontal extent.

difficult to detect seismically that it cannot at present be evaluated. Hess (1955) proposed that ultrabasic rock in the mantle may locally be converted to serpentine.

The third solution, strongly favored by the seismic evidence (Byerly, 1938; 1939; Gutenberg, 1943; Richter, 1955, p. 194), may be visualized as a "root" of average crustal material extending downward in mantle rock of assumed specific gravity 3.27. The required minimum thickness of root is then roughly 200/42(3.27 − 2.84) or 11 km, thinning east of the Virginia Range to roughly 170/42(3.27 − 2.84) or 9 km. Depending on the width of the root and the thickness of the crust, the root might be considerably thicker than 11 km and still produce the same gravity anomaly.

Nothing in the gravity data alone favors any one of these solutions over the others. The abundance of granitic rocks in the Sierra Nevada, however, suggests that part of the negative anomaly is due to granite in the crust. The extra thickness of granitic rocks necessary to explain the change in Bouguer anomaly (30 mgals) from the Sierra Nevada to the area east of the Virginia Range amounts to only about 5 km. Hence, 5 km of granite, replacing average crust in the Sierra region and lacking in the Basin Range area to the east, would fully account for the difference in Bouguer anomaly between these areas. The whole gravity picture, then, might be explained as due to an eastward thickening of the crust beneath the higher part of the continent, combined with extra granitic rock in the crust of the Sierra region. No local root beneath the Sierra would be required.[4]

The uncertainty of interpretation emphasizes the need for more detailed seismic measurements in the Sierra Nevada region and particularly for measurements on the eastern limit of the seismically indicated root. Such measurements should contribute to an understanding of the geologic history of the Sierra, which presents many challenging problems. After intense folding and the formation of batholiths in the Mesozoic era, the region seems to have lain relatively dormant until the latter part of the Cenozoic era, when it was uplifted to its present impressive height. If the crust was thickened and a root formed during the Mesozoic folding, why did the great uplift occur many millions of years later? Perhaps the uplift depended not on buoyant elevation by an old root but on expansion and phase changes in the material below the mountains. Of many possible phase changes three may be mentioned: crystalline rock to glass, peridotite to serpentine, and eclogite to basalt. Perhaps seismic tests can be devised to decide these questions.

[4] After this was written, Romney (1957) reported seismic evidence indicating regional thickening of the crust from west to east in the Sierran region. A local root is indicated beneath only the central and southern part of the Sierra Nevada and not beneath the lower northern part discussed in the present report.

Conclusions

Gravity measurements indicate that basin sediments in Washoe Valley are at least 1800 feet thick. In the northern part of Steamboat Valley, the maximum thickness of the sediments is only about 600 feet, and there is no large displacement of the concealed bedrock surface on faults. In the large area of Truckee Meadows, sedimentary fill attains a thickness of 1000 feet or more in several places; between the Virginia Range and the town of Sparks, an 18-mgal anomaly suggests that the fill has a thickness of roughly 2800 feet, greater than the topographic relief of the range at this point. In the basin west of Reno, sediments more than 1800 feet thick are indicated by the gravity data, and this thickness is verified by an old drill hole.

Granitic rocks produce negative anomalies relative to the metamorphic rocks. Altered Tertiary volcanic rocks do not produce prominent negative anomalies, and this suggests that intense leaching is only a surface feature.

The Bouguer gravity anomaly in the ranges is not significantly different from that in sediment-free parts of the basins, as though the mountains were not isostatically compensated but were excess loads piled on an unyielding surface. The gravity measurements, however, cannot resolve small differences resulting from masses that are deep compared to the width of the ranges.

On a large scale, the Bouguer anomaly from San Francisco eastward generally shows an inverse relationship to elevation but is a minimum along the east side of the Carson Range rather than at the main summit of the Sierra. The gravity profile may reflect a thickening of the crust eastward beneath the higher part of the continent, combined with a smaller effect from greater-than-average amounts of granitic rock in the crust of the Sierran region. The answer will depend on more precise and more detailed seismic measurements.

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STRUCTURE OF THE CUMBERLAND PLATEAU, TENNESSEE

BY CHARLES W. WILSON, JR., AND RICHARD G. STEARNS

ABSTRACT

Present structure of the Cumberland Plateau results from two components: (1) regional eastward dip of about 25 feet per mile off the Cincinnati arch; and (2) Pine Mountain, Cumberland Plateau, and Sequatchie Valley bedding thrusts, which strike northeastward and have roots in the Valley and Ridge province to the southeast.

Gentle eastward dip off the flank of the Cincinnati arch is the only structure affecting the undisturbed portion of the plateau, and is also the primary component of structure in the faulted area. Bedding thrusts do not disturb the position and attitude of strata within the thrust sheets. Locally, however, where the faults cut across bedding, superficial anticlines are formed. The structural concordance between the undisturbed areas and areas of bedding thrusts supports the "no-basement" interpretation for the essential regional structure.

Movement of the Cumberland Plateau overthrust sheet over steep portions of its generally flat fault plane caused superficial flat-topped, steep-sided anticlines similar to the Powell Valley anticline of the Pine Mountain overthrust. Abrupt ends of such anticlines mark places where buried steep thrusts become buried strike-slip faults.

The oldest thrust is the northwest-dipping Whetstone Mountain fault; next is the Cumberland Plateau overthrust; the youngest thrusts are the Sequatchie Valley thrust and some faults of the Valley and Ridge province.

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INTRODUCTION

The Cumberland Plateau extends across Tennessee from Kentucky and Virginia on the north to Georgia and Alabama on the south. Regional physiographic and structural features are shown on Figure 1. Although the dominant structural pattern of the plateau is a gentle dip to the east off the Cincinnati arch, this slope is broken by the surface trace of the Pine Mountain and Cumberland Plateau overthrusts, the Sequatchie Valley anticline, and by an abrupt upturn along the eastern escarpment of the plateau (Fig. 2).

The Cumberland Plateau is underlain by Pennsylvanian rocks. Mississippian rocks crop out on lower slopes of the western escarpment, but beds as old as the Knox dolomite of Cambro-Ordovician age are exposed in Sequatchie Valley, and Cambrian strata are brought up by thrusting along the eastern escarpment.

This report is primarily concerned with the configuration of Pennsylvanian rocks. Figure 3 shows the generalized stratigraphic column of the Pennsylvanian. Discussion of the structure which follows is arranged in geographic order by the structural subprovinces shown on Figure 2.

Previous Investigations

Safford (1869), the first to describe the Cumberland Plateau structure, recognized the Sequatchie Valley anticline, the Pine Mountain block, and the sharp upturn at the eastern edge of the plateau which marks its boundary with the Valley and Ridge province. He also noted disturbances now known to be along the trace of the Cumberland Plateau overthrust but considered them to be of only local significance.

More detailed descriptions include those of Hayes (1894a; 1894b; 1894c; 1895a; 1895b),

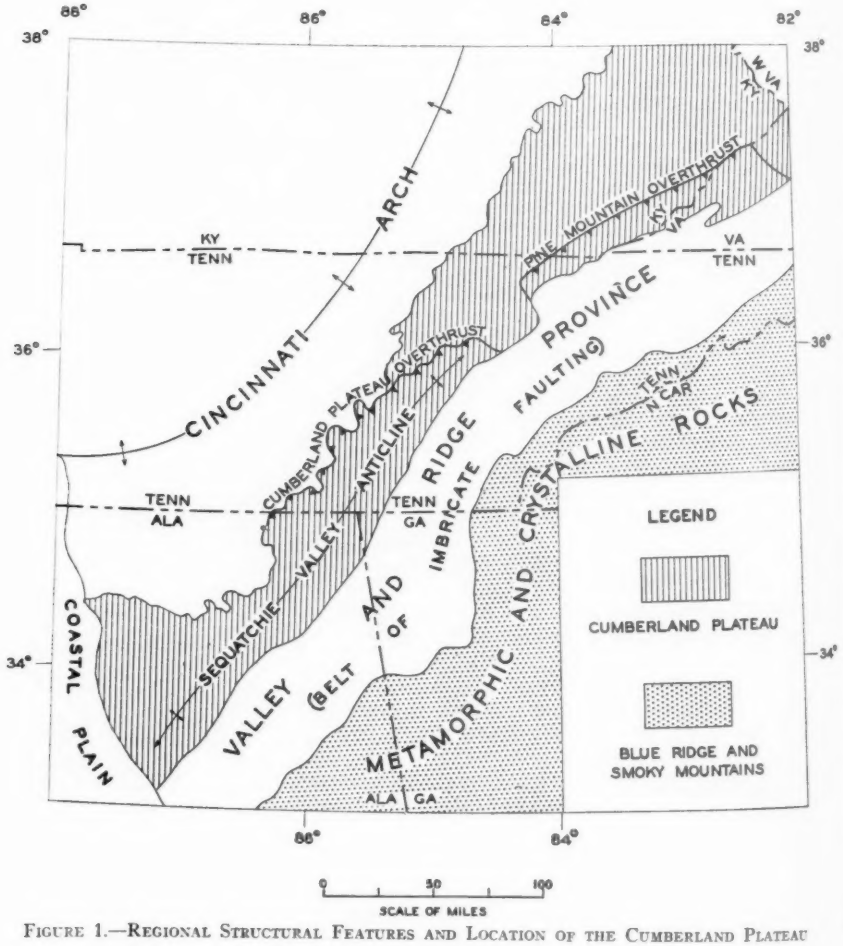

Figure 1.—Regional Structural Features and Location of the Cumberland Plateau

Keith (1896; 1897), Campbell (1899), Butts (1916; 1919; 1927), Butts and Nelson (1925), Glenn (1925), and Nelson (1925a; 1925b).

Wentworth (1921), Butts (1927), and Rich (1934) established that the Pine Mountain block is part of a low-angle thrust sheet which has its roots in the Valley and Ridge province to the southeast. Rich (1934) suggested that the Sequatchie Valley anticline similarly resulted from low-angle thrusting. Rodgers (1950) presented new information supporting this concept and suggested that outlying structures northwest of the Sequatchie Valley anticline are also part of an incipient thrust sheet. His geologic map of East Tennessee (1953a) includes the southeastern boundary of the plateau. Rodgers (1953b) prepared a cross section of the combined Valley and Ridge and Cumberland Plateau which interprets the visible structures as confined to the near-surface rocks and not involving the lower strata or the basement. The present investigation mainly bears out his interpretation.

Miller and Fuller (1954) did detailed work on the Pine Mountain overthrust in Virginia, near the Tennessee line. Stearns (1954; 1955) demonstrated that most of the southern Cumberland Plateau has moved northwestward along the relatively flat Cumberland Plateau overthrust in a manner similar to the Pine Mountain overthrust to the north. Swingle *et al.* (1956) reported that Pennsylvanian beds, previously thought to be restricted to the plateau, lie buried by thrusting in adjacent areas of the Valley and Ridge province. They concluded that the Cumberland Plateau and Valley and Ridge province are not sharply divided but merge structurally. Wilson *et al.* (1956) presented structure maps and a summary of the structural features of the plateau; this information is part of a folio on general Pennsylvanian geology resulting from field work by many members of the Tennessee Division of Geology, including the present writers.

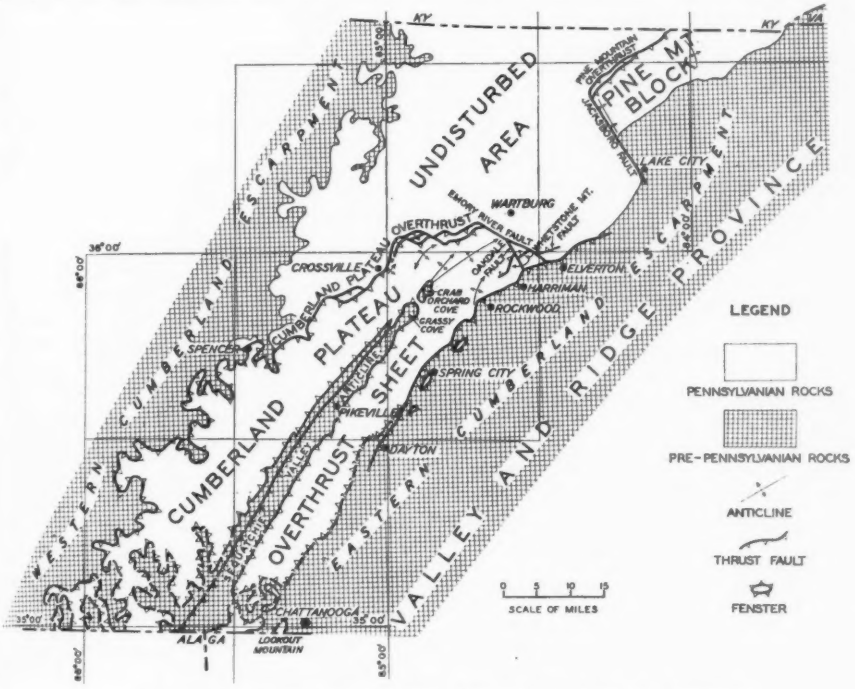

FIGURE 2.—STRUCTURAL FEATURES OF THE CUMBERLAND PLATEAU, TENNESSEE

ACKNOWLEDGMENTS

The writers are indebted for structural information to many members of the Tennessee Division of Geology who participated in field

mapping of the Cumberland Plateau. Chief among these are John W. Jewell, Edward T. Luther, Robert M. Mitchum, Jr., and Robert L. Wilson.

Special thanks are due to W. D. Hardeman, State Geologist of Tennessee, who has maintained active support and interest in the geology of this area. He spent much time in the field and contributed many ideas and helpful suggestions.

The writers are grateful to George D. Swingle of the University of Tennessee and Herbert A. Tiedemann of the Tennessee Division of Geology for information on the eastern Cumberland escarpment and adjacent portions of the Valley and Ridge province.

Robert J. Floyd, Senior Geologist of the Tennessee Division of Geology, edited the manuscript.

Structure of the Cumberland Plateau

Structurally, the Cumberland Plateau in Tennessee is divided into three subprovinces: (1) an area undisturbed by thrust faulting, comprising less than half the province; (2) the Pine Mountain overthrust sheet; and (3) the Cumberland Plateau overthrust sheet (*see* Fig. 2). Both the Pine Mountain and Cumberland Plateau overthrusts are thin sheets that moved primarily along extensive bedding thrusts. The thin Cumberland Plateau overthrust sheet is interrupted by the deeper-seated prominent Sequatchie Valley anticline. The southeastern boundary of the plateau is a prominent escarpment that locally is dominated by Valley and Ridge structure.

Figure 4 is a structural contour map of the Cumberland Plateau drawn on the Rockcastle conglomerate as a datum. This formation has been removed by erosion over much of the southern part of the plateau, where the configuration of the older Sewanee conglomerate has been projected to the Rockcastle.

Pine Mountain Overthrust

General Statement

The Pine Mountain overthrust bounds a rectangular thrust sheet extending 125 miles from northeast Tennessee into Kentucky and Virginia. The northwest portion of the overthrust sheet is a broad syncline, underlain by Pennsylvanian rocks (herein called the Pine Mountain block), that is part of the Cumberland Plateau; the southeast part is the broad, flat-topped Powell Valley anticline in the Valley and Ridge province (Fig. 1); Cambrian and Ordovician rocks are exposed at the surface of the anticline. The southwestern end of the thrust sheet moved about 10 miles. Displacement diminishes northeastward; near the Tennessee border in Virginia, the movement was only 5.8 miles (Miller and Fuller, 1954). Butts (1927), Rich (1934), and Miller and Fuller (1954) have discussed the structure and mechanics of this block in detail. Figure 5 shows a cross section of the Pine Mountain overthrust near its southwest end. The location of this cross section is plotted on Figure 4.

Pine Mountain Block

The Pine Mountain block refers only to that part of the larger thrust sheet that involves Pennsylvanian rocks. It is separated from the main body of the plateau by the Pine Mountain and Jacksboro faults. The Pine Mountain block is surrounded by prominent strike ridges. It is bounded on the southeast by a steep northwest dip along the eastern Cumberland escarpment. This is the steep northwest flank of the Powell Valley anticline. On the northwest is Pine Mountain, which reflects a 10- to 35-degree southeast dip of resistant beds on the hanging wall of the Pine Mountain fault. To the southwest, on the northeast side of the Jacksboro fault, is a prominent anticlinal ridge. Along all the strike ridges at the periphery of the block, dip flattens abruptly to merge with the relatively flat floor of the syncline.

Within the block is a rectangular structural basin with nearly flat beds throughout most of its extent. It is, however, underlain by a bedding thrust, which has its roots in the Valley and Ridge province to the southeast. Along this thrust Pennsylvanian strata have been moved northwestward about 10 miles. Corresponding Pennsylvanian beds within the block are about 500 feet above their normal level on the adjacent parts of the plateau.

Pine Mountain Thrust Fault

At the trace of the Pine Mountain fault, beds as old as Silurian have been thrust over Pennsylvanian. The fault extends more than 100 miles beyond Tennessee into Kentucky and Virginia. Northwest of the Pine Mountain thrust in Tennessee is a subsidiary fault slice which consists of intensely deformed beds of the Gizzard and Crab Orchard Mountains

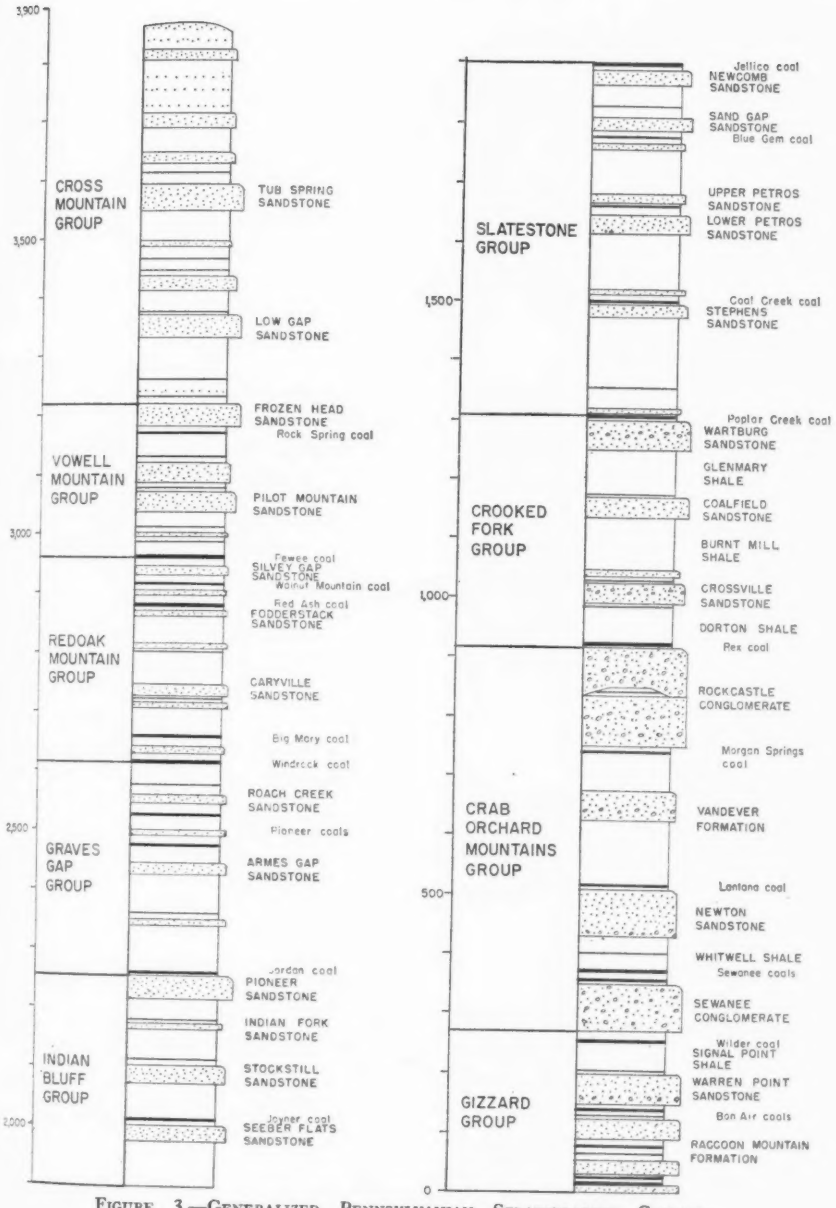

FIGURE 3.—GENERALIZED PENNSYLVANIAN STRATIGRAPHIC COLUMN FOR THE CUMBERLAND PLATEAU, TENNESSEE

groups. The gentle dip of this fault is shown by an outlying klippe known as Hells Point Ridge, where brecciated older sandstones occur above flat-lying younger coal-bearing beds (*see* Fig. 5).

Jacksboro Cross Fault

The southwestern boundary of the Pine Mountain block is the Jacksboro cross fault.

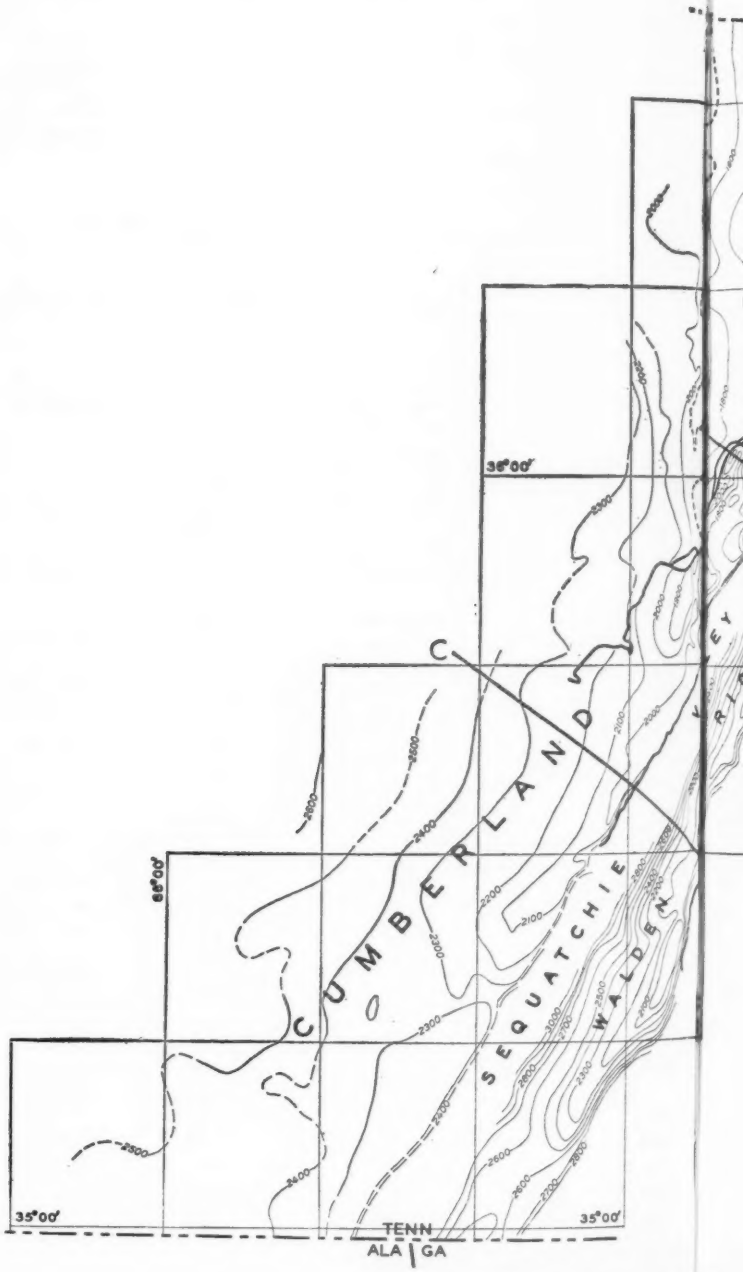

FIGURE 4.—STRUCTURE CONTOUR MAP CUMB
Lines A-A', B-B', and C-C' locate the cross

NTOUR MAP CUMBERLAND PLATEAU, TENNESSEE
C' locate the cross sections, Figures 5 and 6.

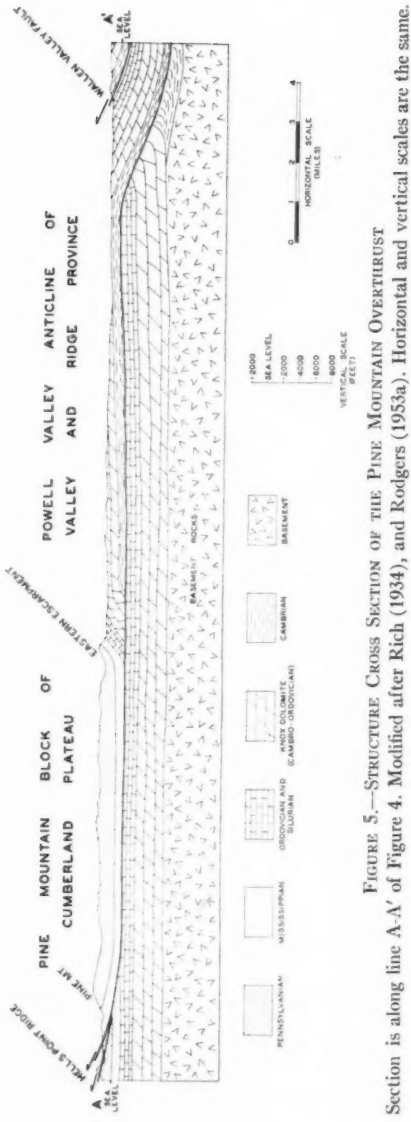

FIGURE 5.—STRUCTURE CROSS SECTION OF THE PINE MOUNTAIN OVERTHRUST

Section is along line A-A' of Figure 4. Modified after Rich (1934), and Rodgers (1953a). Horizontal and vertical scales are the same.

This fault is nearly vertical at the surface, but northeastward, in the subsurface, it flattens abruptly and merges with the Pine Mountain thrust. This fault moved laterally approximately 10 miles, but its vertical movement was only about 500 feet. The "crowding" accompanying lateral movement resulted in the tight anticline along the northeast side of the fault (Fig. 2). Here, the deformed beds of the Crab Orchard Mountains and Gizzard groups are faulted against nearly flat Slatestone beds (Fig. 3). Southeast of the Pine Mountain block, the Jacksboro fault extends along the plateau edge as far as Lake City (Fig. 2).

CUMBERLAND PLATEAU OVERTHRUST

General Relationships

The Cumberland Plateau overthrust limits a thrust sheet, similar to the Pine Mountain block, which includes the entire southern half of the plateau. The northern limit of the overthrust sheet is a complex series of faults, which crosses the plateau from Elverton on the east to Spencer on the west (Fig. 2). The sheet continues southwestward as a bedding thrust along the western escarpment face at least as far as the southern border of Tennessee. It also crops out as a bedding fault around Sequatchie Valley, Grassy Cove, and Crab Orchard Cove, making fensters of these valleys.

Regionally, movement of the Cumberland Plateau overthrust sheet appears to have been rotational, diminishing southwestward, mirroring the rotational movement of the Pine Mountain overthrust sheet which diminishes northeastward (Butts, 1927). This thrust sheet is known to extend nearly to the Alabama border; the writers believe that it continues southward into Alabama, although the movement may have been slight.

Cumberland Plateau Overthrust Sheet

Like the Pine Mountain block, this southern part of the plateau is in general underlain by gently dipping strata (Fig. 4). Also, like the Pine Mountain block, this extensive area of nearly flat beds is completely underlain by bedding thrusts of the Cumberland Plateau overthrust. Pennsylvanian beds have been detached from underlying rocks and thrust as far as 3 miles northwestward. A major interruption of this area is the Sequatchie Valley anticline, which strikes northeast throughout the entire length of the thrust sheet.

Between the Sequatchie Valley anticline and the eastern escarpment is an asymmetrical syncline, with the axis close to the eastern escarpment, except in the southernmost part

near Chattanooga, where it is more symmetrical. This syncline includes a series of elongate gentle basins; it generally pitches northeastward, and the lowest structural position within the Cumberland Plateau overthrust sheet occurs in this syncline east of the end of the Sequatchie Valley anticline. Here, the syncline axis is about 1400 feet lower than it is at the Georgia-Tennessee border. Similarly, a syncline, the axis of which lies just west of the Sequatchie Valley anticline, pitches northeastward from the Alabama-Tennessee border to the Emory River fault where the strata are about 1300 feet lower. (*See* Fig. 4.)

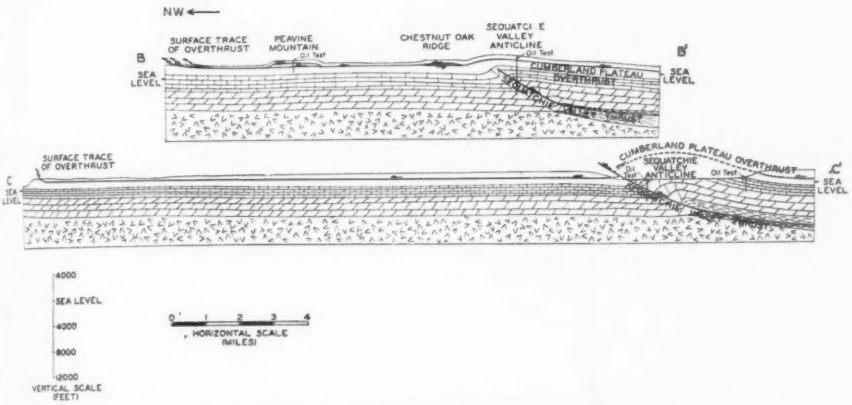

FIGURE 6.—STRUCTURE CROSS SECTIONS OF THE CUMBERLAND PLATEAU OVERTHRUST AND THE SEQUATCHIE VALLEY ANTICLINE
Sections along lines B-B′ and C-C′ of Figure 4. Horizontal and vertical scales are the same.

Outcropping Faults of the Overthrust System

From Elverton to Spencer, across the plateau, the overthrust system crops out as a series of echelon thrusts joined by vertical cross faults. At the surface these thrusts dip more than 30° southeast, but the cross faults are vertical. Along the western escarpment south of Spencer, and around Sequatchie Valley, Grassy Cove, and Crab Orchard Cove, the system crops out as bedding thrusts.

The Emory River fault, first described by Jillson (1923), is the northeast limit of the Cumberland Plateau overthrust. It is the counterpart of the Jacksboro fault of the Pine Mountain overthrust. This fault moved laterally about a mile. The Emory River fault is overridden by Valley and Ridge thrusts at the escarpment. The Oakdale fault is a branch of the Emory River fault that intersects the eastern escarpment at Harriman. This fault is also overridden by a later thrust of the Valley and Ridge province (G. D. Swingle, personal communication). Other individual thrust and cross faults of this system have been named and described in detail by Stearns (1954).

Thrusts and cross faults are portions of the same complex fault plane, because they do not offset one another, but turn and merge. Thrusts in this belt are easily traced owing to the prominent ridges formed from tilted resistant sandstone formations, but cross faults are traced with difficulty because of lack of stratigraphic offset. Southwest of Spencer and around Sequatchie Valley, where the fault is a bedding thrust, the only effect is contortion and shearing of shales and coal in zones a few feet thick.

The Cumberland Plateau overthrust extends through a sequence of Pennsylvanian sandstones and conglomerates interlayered with incompetent coal-bearing shales. As a bedding thrust it follows the incompetent shales over extensive areas (Fig. 6); as a thrust or a cross fault it breaks through competent sandstones along narrow zones.

The configuration of the fault plane is known from exposures in fensters (Sequatchie Valley, Grassy Cove, and Crab Orchard Cove) and along the fault trace to the north and west. In these exposures the fault is a bedding thrust in various incompetent shaly zones. Its position rises stratigraphically to the northwest in the direction of fault movement, and

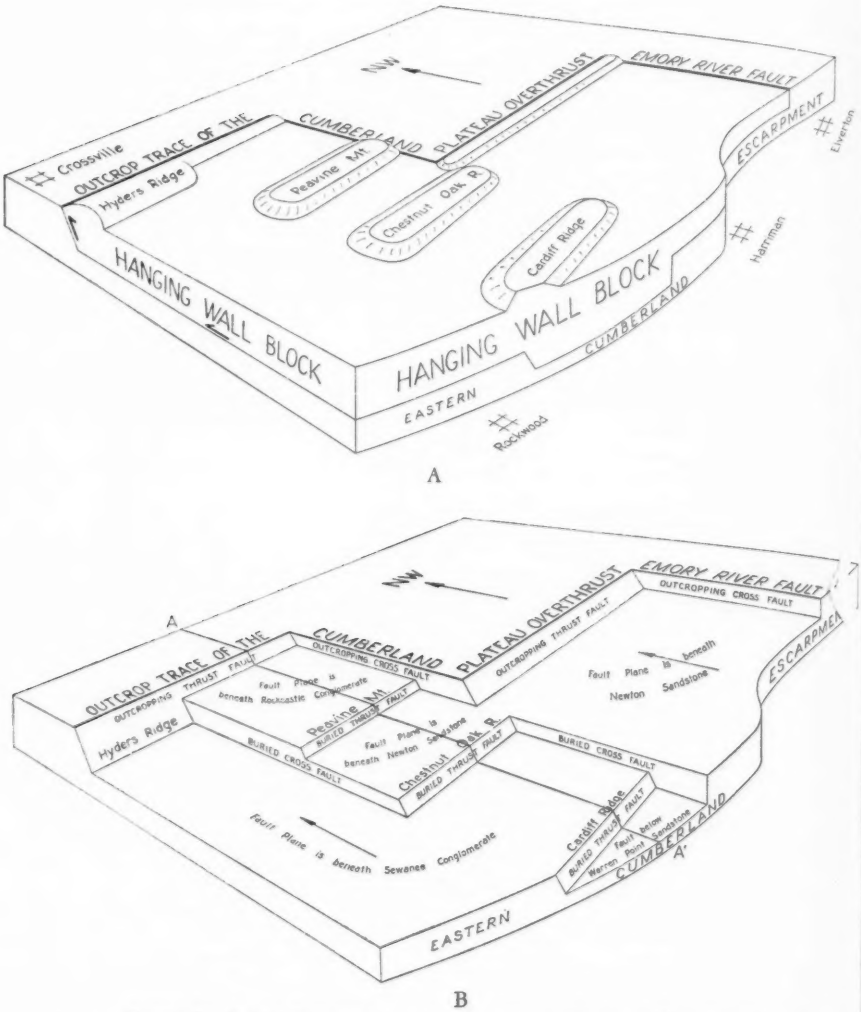

FIGURE 7.—BLOCK DIAGRAMS OF THE CUMBERLAND PLATEAU OVERTHRUST
A shows the location of the surficial anticlines. *B* shows the configuration of the fault plane. The thrust sheet, or hanging wall block, is removed in *B*, exposing the footwall.

to the northeast across the direction of movement.

Surficial Anticlines Caused By Thrusting

Drilling has revealed that Peavine Mountain, surficially an anticline, is underlain at a shallow depth by flat beds. Uplift is due to subsurface duplication of beds where the Cumberland Plateau overthrust breaks through the competent Newton sandstone. The main fault crops out 5 miles northwest of Peavine Mountain, where it thrusts Rockcastle conglomerate over Crossville sandstone. Peavine Mountain has a broad, flat top and steep sides similar to the Powell Valley anticline of the Pine Mountain overthrust (Fig. 5). This square shape in cross section is characteristic of anticlines formed over a steep segment of a buried low-angle thrust, along which there has been considerable movement (Rich, 1934).

The subsurface configuration of the Cumber-

land Plateau overthrust can be inferred from the fault exposures and from the position of Peavine Mountain and similar anticlines. Buried steep thrusts are marked by the anticlines, but buried cross faults leave no surface as shown on Figure 6. Had there been any arching of this anticline before the overthrusting occurred, the overthrust would have broken through to the surface along the Sequatchie Valley anticline.

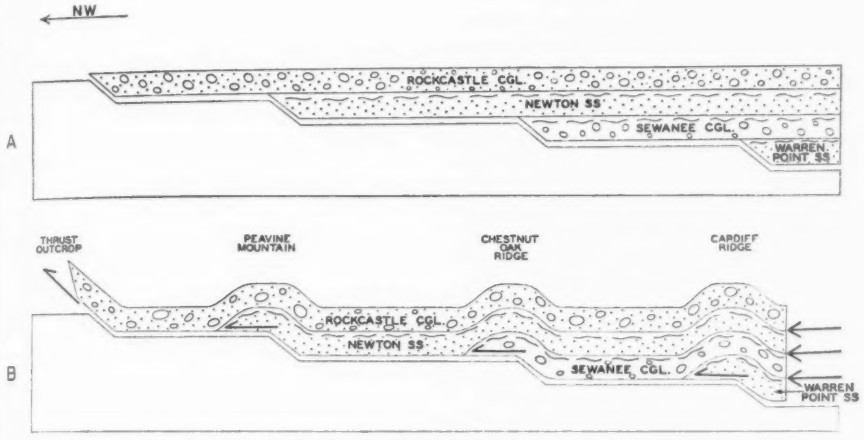

FIGURE 8.—MECHANICS OF SUPERFICIAL ANTICLINE FORMATION
A is before thrusting, *B* is after thrusting. The location of this cross section is shown on Figure 7*B* (section A-A').

indication. The 3 miles of movement responsible for Peavine Mountain structure cannot end abruptly, as does the anticline itself, without a cross fault. The end of the anticline, which marks the end of a buried thrust, also marks the junction of this buried thrust with a buried cross fault.

Figure 7*A* is a block diagram showing the relative positions of the thrust outcrop and the surficial anticlines. On Figure 7*B* the thrust plate is removed to show the buried fault-plane configuration of the Cumberland Plateau overthrust interpreted from the position of the anticlines and from exposures of the fault. Figure 8 is a diagrammatic cross section showing the mechanics of formation of three anticlines. Structure of the eastern escarpment and the northeast end of the Sequatchie Valley anticline is omitted on Figures 7 and 8 to simplify the relationships of the thrust plate.

Relationships of the Cumberland Plateau Overthrust to Other Structures

The Cumberland Plateau overthrust predates the Sequatchie Valley anticline, because it is folded by the Sequatchie Valley anticline and is offset by the Sequatchie Valley thrust.

Prior to movement of the Cumberland Plateau overthrust, the Whetstone Mountain fault was formed parallel to Valley and Ridge structure (Fig. 2). It is exceptional because the fault plane dips northwest. This structure has been offset approximately 1 mile by the Emory River fault of the Cumberland Plateau overthrust and has probably been offset a lesser distance by the Oakdale fault (Fig. 2).

The northwest-dipping Whetstone Mountain fault is the oldest structure described; the Cumberland Plateau overthrust is the next oldest; and the Sequatchie Valley anticline and at least some of the Valley and Ridge thrusts are the youngest.

Another low-angle overthrust, named the Ozone thrust (Stearns, 1954), occurs within the Cumberland Plateau overthrust mass and crops out east of Crab Orchard Cove, where the Newton sandstone is repeated. The fault is not shown on the illustrations, because its extent is unknown. The Ozone fault is the same type of structure as the main overthrust, and future work may reveal other bedding thrusts. This fault probably is approximately the same age as the Cumberland Plateau overthrust.

Sequatchie Valley Anticline

This long, straight anticline extends 200 miles from Blount County, Alabama, northeastward into Morgan County, Tennessee (Fig. 1). The southeastern flank dips gently, but the northwestern flank is steep and broken by the Sequatchie Valley thrust throughout most of the length of Sequatchie Valley. Along this thrust, beds as old as Cambro-Ordovician are carried over Pennsylvanian strata. The anticline pitches northeastward and terminates at the Emory River fault (Fig. 2).

From the Alabama-Tennessee state line to a point 10 miles north of Pikeville, resistant Pennsylvanian strata along the crest of this anticline have been removed by erosion, exposing less-resistant carbonate rocks into which the prominent Sequatchie Valley has been cut (Fig. 2). Northeast of the head of Sequatchie Valley, Pennsylvanian strata have been breached locally, and erosion has cut into the underlying Mississippian beds to form two valleys, Grassy Cove to the south and Crab Orchard Cove to the north (Fig. 2). Still farther northeast, the Crab Orchard Mountains mark the unbreached portion of the diminishing Sequatchie Valley anticline.

The Sequatchie Valley fault, as a surface feature, ends near the head of the main valley. The Shell Oil Company's Peterson No. 1 well adjacent to Crab Orchard Cove has revealed that 14 miles beyond the northeast end of the surface fault the anticline is faulted at depth (B-B′, Fig. 6). In this well, about 1750 feet of Middle Ordovician beds are repeated. This amount of repetition is significant, because it accounts for the entire surface structural relief of the anticline in this area.

The Cumberland Plateau overthrust is folded by the Sequatchie Valley anticline (as shown in sections B-B′ and C-C′, Fig. 6) and is offset by the Sequatchie Valley fault (section C-C′, Fig. 6).

The isolated position, straightness, and length of this anticline have long been regarded as proof that basement rocks are involved in its structure at depth. Rich (1934), however, proposed that the anticline results from low-angle thrusting through Cambrian shales with roots far to the east in the Valley and Ridge province, and that the underlying basement is undisturbed. Rodgers (1950) emphasized this and pointed to cross faults at either end of the anticline as proof. However, the Emory River cross fault at the northeast end of the anticline is now known to be part of the superficial older Cumberland Plateau overthrust rather than part of the Sequatchie Valley structure. The present writers agree with Rodgers and Rich and believe that the Sequatchie Valley fault is a low-angle overthrust in the subsurface like the Pine Mountain and Cumberland Plateau overthrusts, as shown on Figure 6.

Other writers have postulated that some of the low-angle thrusts of the Valley and Ridge province occur as bedding faults in Rome and Conasauga shales. During the Middle Ordovician the future site of Sequatchie Valley occupied a foreland shelf position and was not in the Appalachian geosyncline (Wilson, 1949, p. 323). It is believed likely that the Rome and lower Conasauga shales, which were restricted to the geosyncline, overlapped against the foreland near the present position of Sequatchie Valley. This could account for the localization of the Sequatchie Valley structure, the bedding fault being deflected upward where the zone of easy slippage ends. (*See* Fig. 6.)

Eastern Cumberland Escarpment

A prominent ridge marks the entire southeastern limit of the Cumberland Plateau. The eastern escarpment on the Pine Mountain overthrust is an expression of steeply dipping Pennsylvanian beds forming the northwest limb of the Powell Valley anticline (Fig. 5). The eastern escarpment southeast of the undisturbed area (Fig. 2) also turns up sharply, but here Valley and Ridge faulting lies at the base of the escarpment. The structures here are probably due to crowding against Valley and Ridge faults, where the unyielding part of the plateau abuts the valley.

Southwest of Elverton, the escarpment includes the eastern continuation of the Cumberland Plateau overthrust. Here, the eastern escarpment is locally an anticline. The southeast limb of this feature is commonly buried or faulted out by Valley and Ridge thrusts. Fensters between Dayton and Rockwood (Fig. 2) show that Pennsylvanian beds extend farther southeastward beneath Valley and Ridge thrusts.

Undisturbed Area of the Cumberland Plateau

West of the Pine Mountain block and north of the Cumberland Plateau overthrust the plateau is undisturbed by bedding thrusts (Fig. 2). Between the two thrust sheets the undisturbed area extends to the edge of the Valley

and Ridge province, and beds are upturned abruptly only at the eastern escarpment. Although this block is separated from the rest of the plateau by faulting at the surface, it is continuous at depth with beds beneath the bedding thrusts of the Cumberland Plateau overthrust and the Pine Mountain block. Also, it is believed to be continuous at depth with the Pennsylvanian fensters between Dayton and Rockwood in adjacent areas of the Valley and Ridge province.

On Figure 4 contours within the undisturbed area strike north, discordant to the northeast strike of the overthrust features and Valley and Ridge structure. This pattern also strongly influences the structure of the disturbed area in the southern Cumberland Plateau, because it continues beneath the thrusts. The southern Cumberland Plateau has a complex configuration owing to local uplift by thrusting, but between the uplifted areas thrusting is by bedding slippage only. These areas of bedding slippage occupy the same structural position they would have if there had been no bedding thrusts.

For example, on Figure 4 the 1900-foot contour line (dashed line) trends northward near the west edge of the plateau in the undisturbed area. South of the undisturbed area, it crosses synclines in places which line up as a continuation of the north trend in the undisturbed area. Other contours may be similarly projected. Also, in the valley, beds in the fensters are at approximately the same elevation as they would be without bedding thrusts. This concordance between the structure of the undisturbed and disturbed areas of the Cumberland Plateau and adjacent portions of the valley is additional verification of the interpretation of the structure of the region that considers that basement rocks are not disturbed.

The gentle east dip off the flank of the Cincinnati arch, which is the only structure affecting the undisturbed area, is also the most important component of structure in the faulted area. Superficial upon this pattern in the southern Cumberland Plateau and Pine Mountain block are the anticlines and fractures due to bedding thrusts.

Summary and Conclusions

Present configuration of Cumberland Plateau strata results from two major structural components—the eastward regional dip off the flank of the Cincinnati arch, and bedding thrusts that have roots in the Valley and Ridge province southeast of the plateau. These outlying bedding thrusts are the Pine Mountain overthrust, Cumberland Plateau overthrust, and Sequatchie Valley thrust. The area they underlie includes more than half the plateau. Even though thrusting is extensive, the whole plateau is underlain by nearly flat beds, which in most places dip about 25 feet per mile eastward off the Cincinnati arch. The only surface indications of the thrusts are dip and contortion in narrow zones adjacent to the fault trace, or anticlines such as Peavine Mountain, Chestnut Oak Ridge, and Cardiff Ridge.

The rectangular Pine Mountain block of northeast Tennessee, Kentucky, and Virginia is part of an overthrust sheet moved 10 miles northwestward on a bedding thrust in the Chattanooga shale and underlying Silurian beds. To the south the similar Cumberland Plateau overthrust sheet includes the entire width of the plateau. This thrust sheet is much thinner and near its northeast end has a maximum of 3 miles of displacement within the lower beds of the Pennsylvanian. Unlike the Pine Mountain overthrust sheet, it is not a rectangular thrust block. Its northern trace is an echelon system of thrusts and cross faults. It crops out as a bedding thrust on the face of the western escarpment. Sequatchie Valley, Grassy Cove, and Crab Orchard Cove are fensters. The Cumberland Plateau overthrust sheet extends to the Alabama border, where its amount of movement may be slight.

Movement of the Cumberland Plateau overthrust sheet over a steep portion of the fault plane is responsible for superficial anticlines, such as Peavine Mountain. These are flat-topped and steep-sided like the Powell Valley anticline of the Pine Mountain overthrust. They mark steep portions of the buried thrust where the fault breaks through competent formations in the subsurface, and the abrupt ends of such anticlines mark loci where buried thrusts turn and become buried strike-slip faults.

The Sequatchie Valley anticline is believed to be the result of a mile or so of low-angle thrusting, perhaps localized in its present position, because Cambrian shales pinch out there. The Sequatchie Valley anticline is faulted at the surface throughout most of its length, and drilling has revealed that the fault continues in the subsurface at least 14 miles beyond the end of the surface outcrop and probably to the end of the anticline.

The oldest structure is the northwest-dipping Whetstone Mountain thrust, which is offset by the northeastern cross faults of the Cumberland Plateau overthrust. Next in age is the Cumberland Plateau overthrust; this pre-dates the Sequatchie Valley anticline, because it is folded along with this anticline and offset by the Sequatchie Valley thrust. The Cumberland Plateau overthrust also pre-dates some of the faults of the Valley and Ridge province, because Valley and Ridge faults override the northeastern cross faults of this system.

Over most of the area underlain by bedding thrusts, the structure is as it would be if there were no thrusts. This concordance between the structure of the undisturbed and disturbed portions of the Cumberland Plateau is additional verification of the structural interpretation that considers the basement to be undisturbed. The easterly dip of 25 feet per mile off the flank of the Cincinnati arch is the only structural component affecting the undisturbed area; it is also the primary component of structure in the faulted area. Superficial upon this basic pattern in the southern Cumberland Plateau and Pine Mountain block are the complications of anticlines and fractures due to bedding thrusts.

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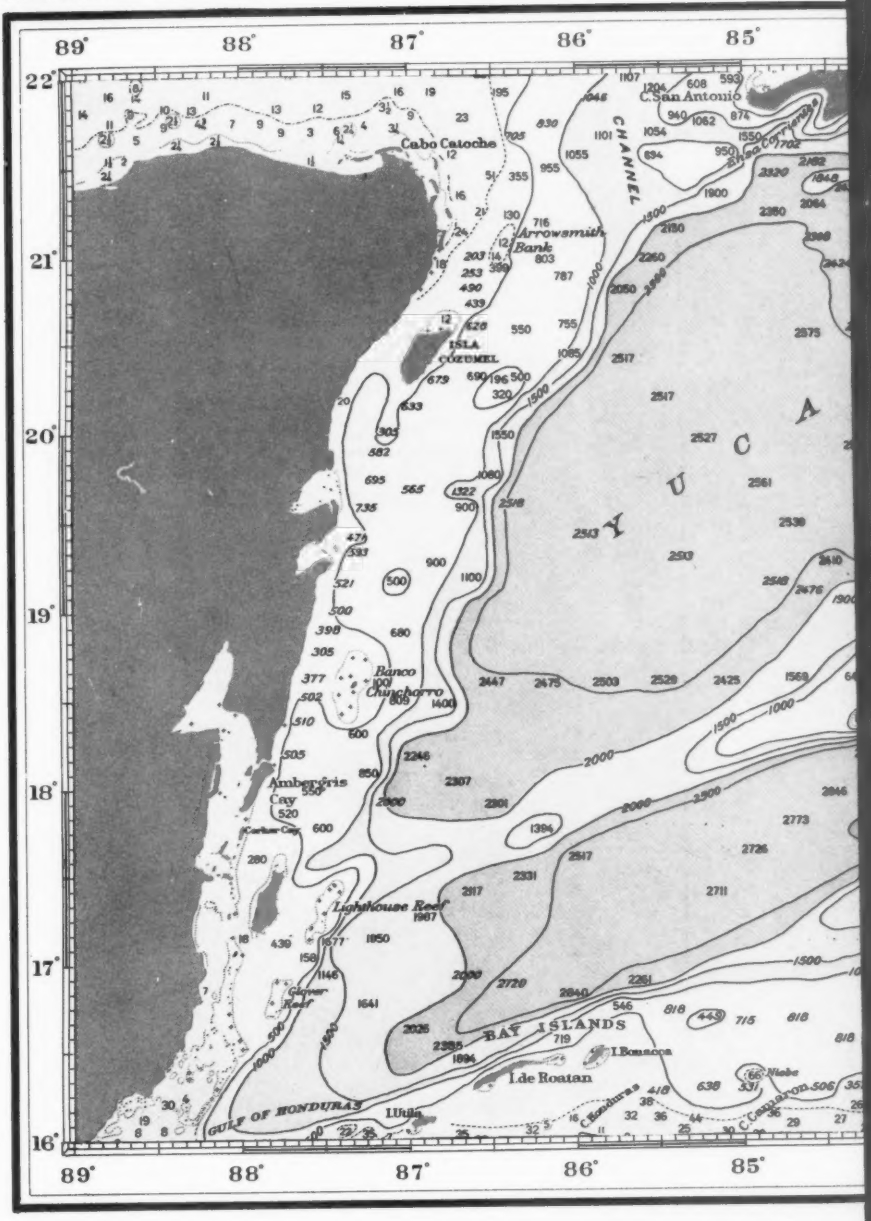
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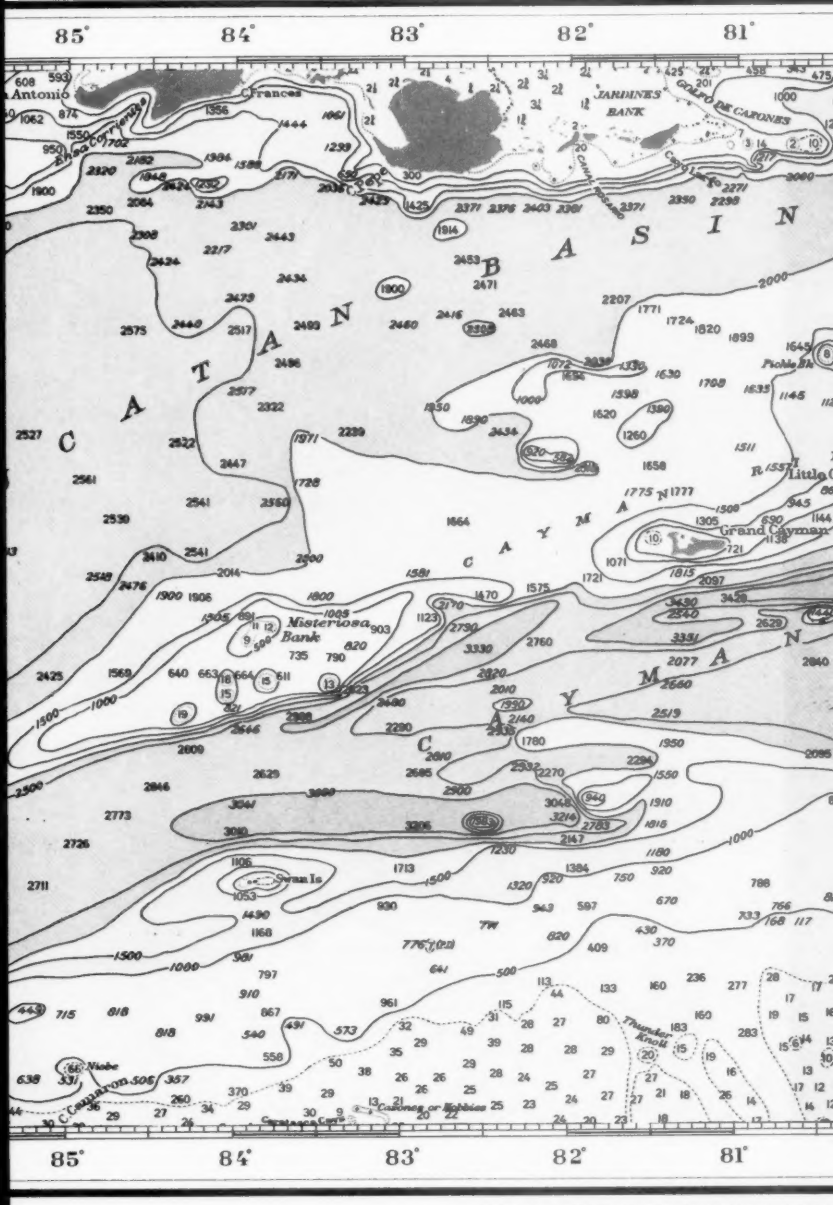

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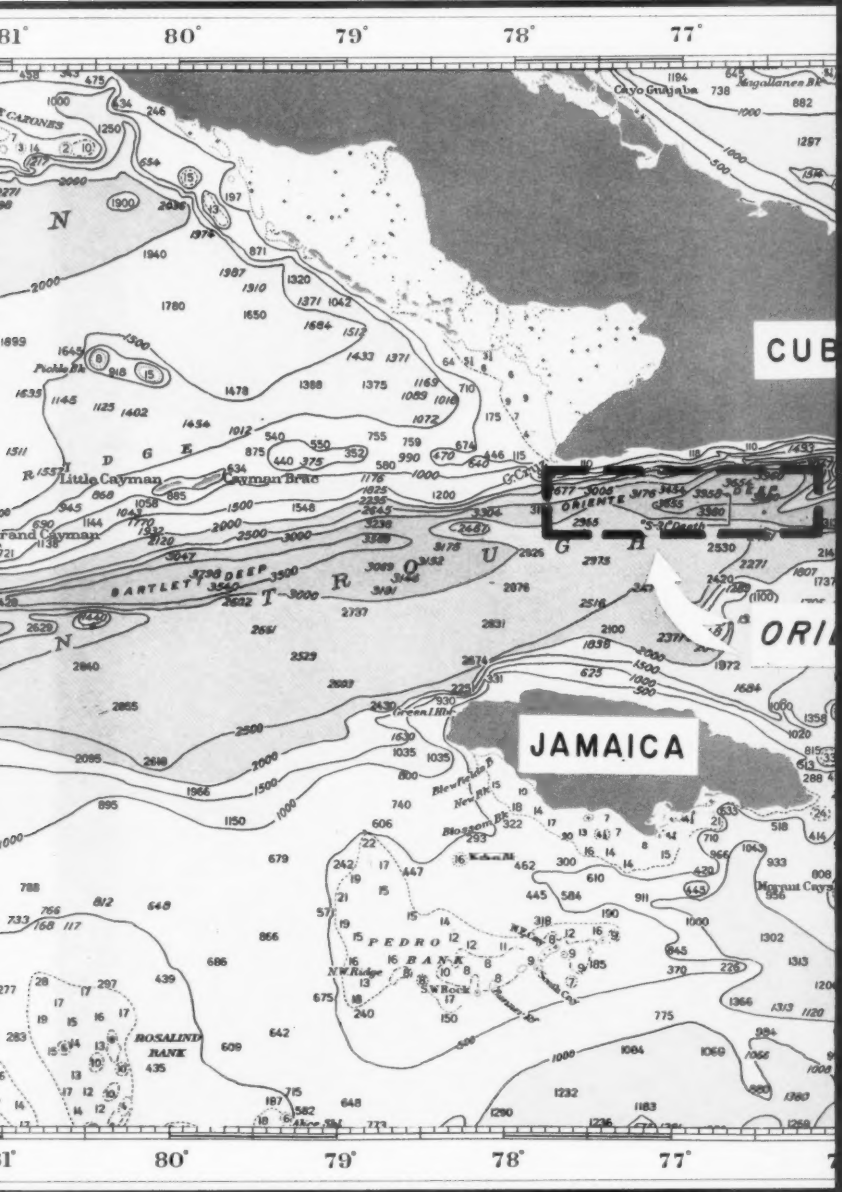

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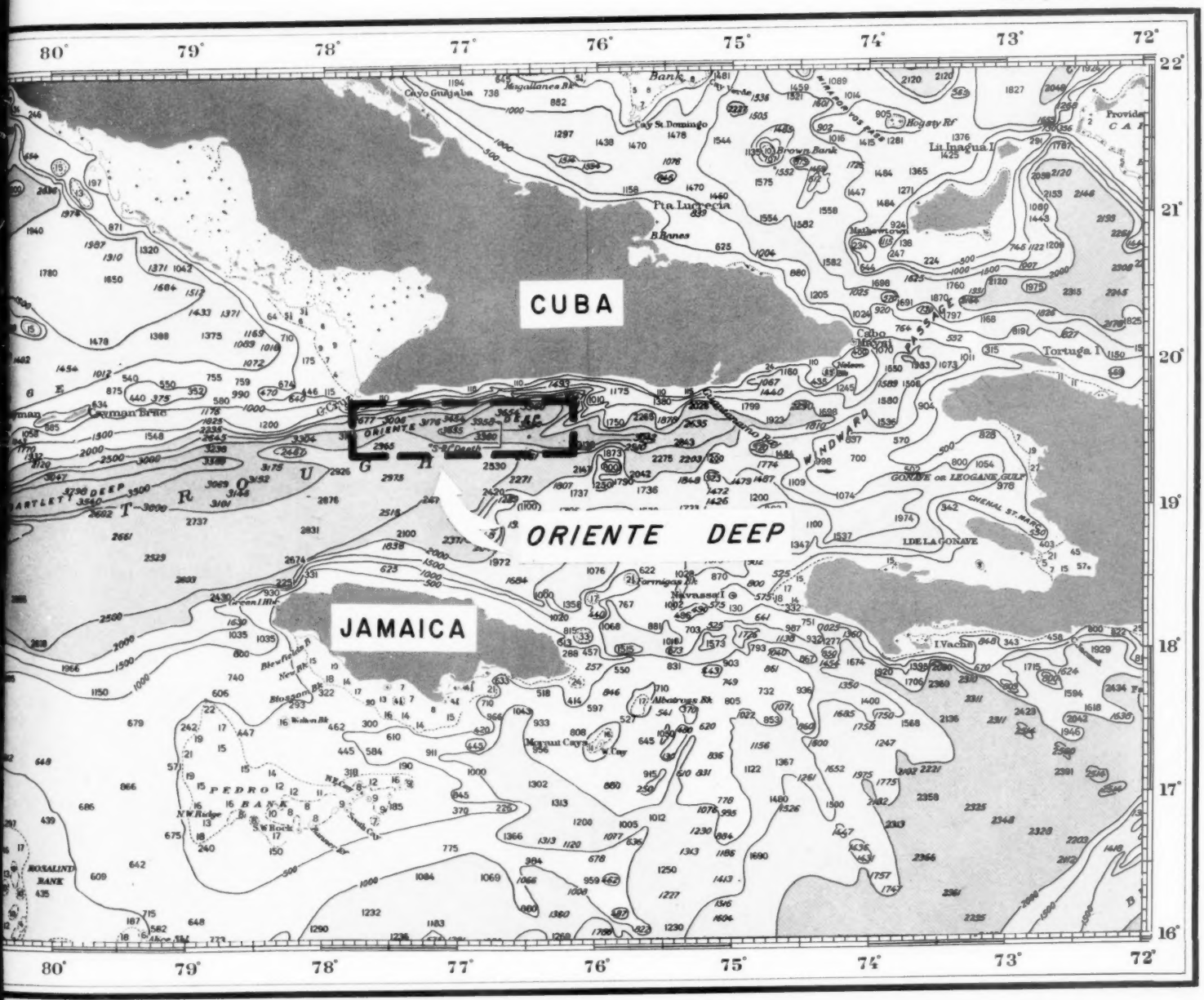

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RECONNAISSANCE SURVEY OF ORIENTE DEEP (CARIBBEAN SEA) WITH A PRECISION ECHO SOUNDER

BY J. B. HERSEY AND MILTON S. RUTSTEIN

ABSTRACT

The submarine slope of the southern coast of Cuba is steep and rugged south of Sierra Maestra, Oriente Province; the average slope in one locality between depths of 1000 and 2000 fathoms is 31°. Two major reversals in slope, at about 2000 fathoms and 2800 fathoms, are taken as evidence of fault zones. Below 2800 fathoms the topography is mountainous in an area parallel to the coast and about 15 miles wide; local relief is several hundred fathoms, but the area deepens to its center where there is a nearly flat plain 3–5 miles wide and about 25 miles long, which also parallels the coast. The plain slopes gradually westward from 3530 to 3542 fathoms with local northward slopes (depths are based on sound velocity of 4800 feet per second); it is the deepest known part of Oriente Deep. The plain is surrounded by hills and intervening gorges which have relatively smooth bottoms grading down to it. Details of the topographic relations indicate that the plain is underlain by unconsolidated sediment brought in by turbidity currents flowing down the gorges. Recent coring in a similar near-by plain supports this view. Seismic evidence, though meager, suggests that the sediment is 1200–1500 feet thick.

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INTRODUCTION

The Oriente Deep lies 15–20 miles south of the southern coast of Oriente Province, Cuba, and is part of the Bartlett or Cayman trough, a deep trench extending from the northwestern tip of the island of Hispaniola more than 1000 miles westward and southward, where it gradually shoals in the Gulf of Honduras. The Bartlett Deep to the west and the Oriente Deep are the deepest parts of this trench (Pl. 1).

During March and April 1956, scientists on the ATLANTIS and BEAR, research ships of the Woods Hole Oceanographic Institution, carried out oceanographic investigations in the Cayman Sea south of Cuba. Depth profiles were recorded across the Oriente Deep with echo sounders fitted with the precision graphic recorders described by Knott and Hersey (1956). On several occasions, especially during the night, when the ships were not otherwise engaged, we were able to make more detailed bathymetric studies there. Although the bathymetric studies had to be subordinated to the other work, we learned much about the geological characteristics of this deep which could be found out only with the aid of depth soundings precise to at least a part in 3000. We also

gained experience in using precisely measured depth profiles both as an aid to navigating the ship at sea and for checking the results of other navigational data after the cruise.

Acknowledgments

The soundings on which this study is based were the work of all the scientists aboard Atlantis and Bear, and the navigational control is the work of the ships' officers. We gratefully acknowledge their considerable contribution to this study. Mr. Charles Innis prepared the figures. We are indebted to the U. S. Hydrographic Office for permission to use a portion of their chart HO-5487 as Plate 1. The cruise during which the soundings were taken was supported under Contract Nonr-1367 with the Office of Naval Research.

Methods

Navigation

In this area such modern electronic aids to navigation as Loran were not available; the tracks of the ship were determined by celestial navigation and dead reckoning. We could make use of cross bearings on mountain peaks on the southern coast of Cuba, but inasmuch as most of the sounding was done at night this method of position fixing was of little use. We might have made long sounding lines starting at known shore points and returning to the coast for closure, but too much time would have been consumed in making observations over a terrain too rugged to map in meaningful detail. From previous visits to the area we knew that the bottom of the Oriente Deep was flat. Hence we elected to study in more detail the shape of the flat area and its relation to the surrounding rugged topography. Since we were generally out of sight of landmarks, the precision of the navigation was by no means comparable to that of the soundings. From the traverses made under the most favorable weather, and when star fixes, sun lines, and dead reckoning agreed well with each other, we outlined the main features of the shape of the flat portion of the deep. On other traverses for which the position control from these data gave results obviously inconsistent with the soundings, we found that in most instances corrections to dead-reckoning data could be applied by noting the times of crossing the edges of the flat area. By thus adjusting the poorer traverses a consistent picture could be mapped. There remain many uncertainties in detail, but we believe that the broad outline is accurate within about 1–1.5 miles.

Soundings

Both ships were fitted with an Edo UQN-1B echo sounder and an 18-inch Woods Hole precision graphic recorder (Knott and Hersey, 1956), which were operated routinely on the 400-, 500-, or 1000-fathom scale. Occasionally, while lying to over the flat area Bear recorded at considerably higher resolution (25- and 50-fathom scales). Time was recorded by interrupting the depth scale every 5 minutes by an electric clock powered by the precision 60 cps power source. Other data such as course and speed of the ship were written on the recorder chart as well as in the deck log.

The power sources generally operated at about 1 part in 3000–5000 less than 60 cps but held a much more nearly constant rate, which was monitored continuously by recording a break-circuit chronometer on the record. All depth scales could be read to a fathom in depth whenever the beginning of the echo was clear. Since this was uniformly the case where the bottom was flat, the echo travel times there are accurate to the equivalent of a fathom in depth at 3000–5000 fathoms. The soundings shown on Plate 6*B* are based on a sound velocity of 4800 feet per second (in conformity with United States custom over the past 10 years), which is considerably too low in this area. We did not take adequate deep hydrographic data to compute an accurate average velocity; approximate average velocity as a function of depth is given in Figure 1 from Matthew's tables (1939) for this area. Although the possibility should not be discounted, significant variation is unlikely in average vertical velocity over an area as small as the Oriente Deep. We assume that small differences in echo travel time are true indicators of differences in depth.

Observations

The adjusted traverses of both ships are shown in Plate 6*A*. The track for each period when soundings were taken was plotted on transparent plastic drafting film with appropriate geographic co-ordinates. Ship tracks over the flat areas (solid lines) are distinguished from those showing relief (dashed lines). Those tracks judged to be based on the best navigation were superimposed on one another. From these the general outline of the flat bottom of the deep could be traced. Other tracks for

which there were several crossings of the edge of the plain were adjusted for best fit by eye by moving their plot over those already established to a best fit with edges of the plain. This procedure adjusts the celestial navigation but not the dead reckoning. Parts of other tracks did not fit the outlines of the flat areas because they indicated that the flat portion was too wide or too narrow. Usually this resulted from not making due allowance for headwinds or cross winds in the original dead reckoning. By adjusting either speed or course made good on these latter tracks over the part for which the position of the plain was already well established, the whole track became much more consistent with the previously established picture. The soundings (Pl. 6*B*) show that the flat portion slopes to the west and in places to the north as well, thus providing approximate position control by its depth pattern. The position control for a few of the traverses was so poor that only by adjusting their soundings to fit could their position be adjusted. In all cases but two this sort of adjustment also made the edges of the flat portion fit satisfactorily. Such adjustments can be influenced by prejudice, but we believe that the cross check provided by comparing both the horizontal extent and depth of the flat area effectively minimizes this danger. Furthermore most of the adjustments were small. The distribution of simple translation adjustments is shown in Figure 2 as a plot of the positions of an arbitrary location before and after adjustment. The center of the diagram is the location before adjustment, and the black dots indicate individual translations. Eight tracks required no adjustment.

The soundings and the outline of the flat portion of the deep are shown in Plate 6*B*;

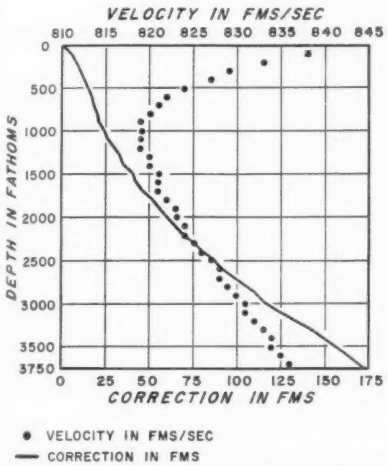


FIGURE 1.—AVERAGE VERTICAL-SOUND VELOCITY After Matthews (1939)

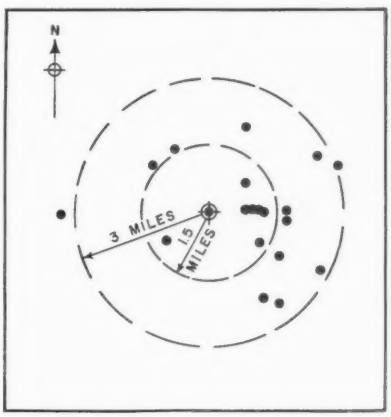


FIGURE 2.—RELATIVE POSITIONS OF AN ARBITRARY LOCATION AFTER ADJUSTMENT OF CELESTIAL NAVIGATION

selected soundings over the surrounding relief are also indicated, but there are far too few soundings over the rough topography to make any but the broadest generalizations about it. In addition to the traverses made to delineate the flat portion a few traverses were made while the ships were en route to or from ports on the southern coast of Cuba. From these is gained a general impression of topography between the island and the deep (Pl. 2). While the ships were lying to over the flat portion a number of high-resolution recordings (50-fathom scale) were made, some of which showed pronounced multiple echoes; unlike multiples often seen over rough terrain, these are unlikely to be side echoes and probably represent penetration into the bottom (Pl. 3). Twice the two ships travelled in company from port; echo soundings were recorded, and frequent radar ranges and bearings were taken between the ships. This operation was intended to extend the usefulness of the single traverse by correlating soundings at the two ships. The records are similar and show qualitatively similar terrain, but the terrain is too rugged to allow detailed correlation of the records taken from ships more than a mile apart. Over the flat portion of the deep the soundings from the two

ships agreed (within 1 fathom); however, once established, this could not aid us further in studying the shape of the deep, since it was seldom possible to survey in company there.

Discussion

The southern coast of Oriente Province, Cuba, trends approximately east-west. Oriente Deep extends parallel to the coast along its western half as far east as Santiago. Farther east the deep terminates in shoaler water. Even the grossest features of the submarine topography are complicated, and the island slope north of Oriente Deep is steep (Pls. 1, 2). There are similar features on other crossings several miles apart. All indicate the steep slope to about 2000 fathoms, then a slight reversal and another somewhat less steep slope to about 2800 fathoms, another reversal, and a still more gradual slope down to the plain. The grade from 2800 fathoms to the plain is irregular and in places mountainous, showing local relief of several hundred fathoms. To the south of the plain there is a zone at least 3–5 miles wide where the topography is similarly complicated.

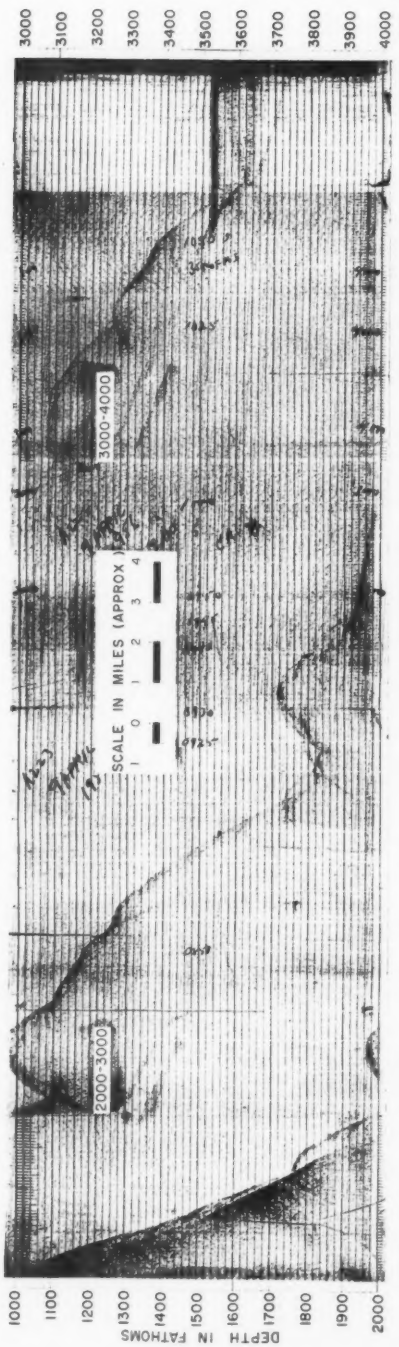
The flat portion of the deep follows the same trend as the coast line and is about 25 miles long and 3–5 miles wide. The sounding lines across the plain indicate that it is as smooth as the echo sounder (3.5 miles away!) is able to determine even when operated on a very high-resolution scale (Pl. 3). Over a considerable part of the plain the returning echo on a high-resolution scale is complex, as in Plate 3. The more or less regular undulation in depth of all the echo sequences is due to the motion of the ship in the surface waves, but the multiple echoes clearly recorded persist with nearly the same, or at least comparable, spacing over a wide area (Pl. 6*B*). We believe that the second group of echoes were reflected from a horizon below the bottom, which occurs widely in the part of the deep where the relief is least. If it is sediment, as we believe it to be, its sound velocity is unlikely to exceed that in the water by much, so that the thickness of sediment to the deeper horizon is about 30 feet. Since a 12-kc sound pulse penetrated the bottom here, the first layer of sediment is probably exceedingly porous; even in shallow water high-frequency sound will not penetrate far into sediments other than unconsolidated silts and muds. A review of several shallow-water studies on sediment penetration by sound is given by Smith and Nichols (1953).

The soundings on the plain deepen from east to west from about 3530 to 3542 fathoms in about 20 miles. In the central portion soundings deepen from about 3534 on the south to 3540 fathoms on the north in 3 miles. In so short a distance in this area the average vertical velocity for soundings probably would not vary appreciably; thus it is reasonable to assume that these slopes are as small as indicated. This is further evidence of a fine, porous sediment.

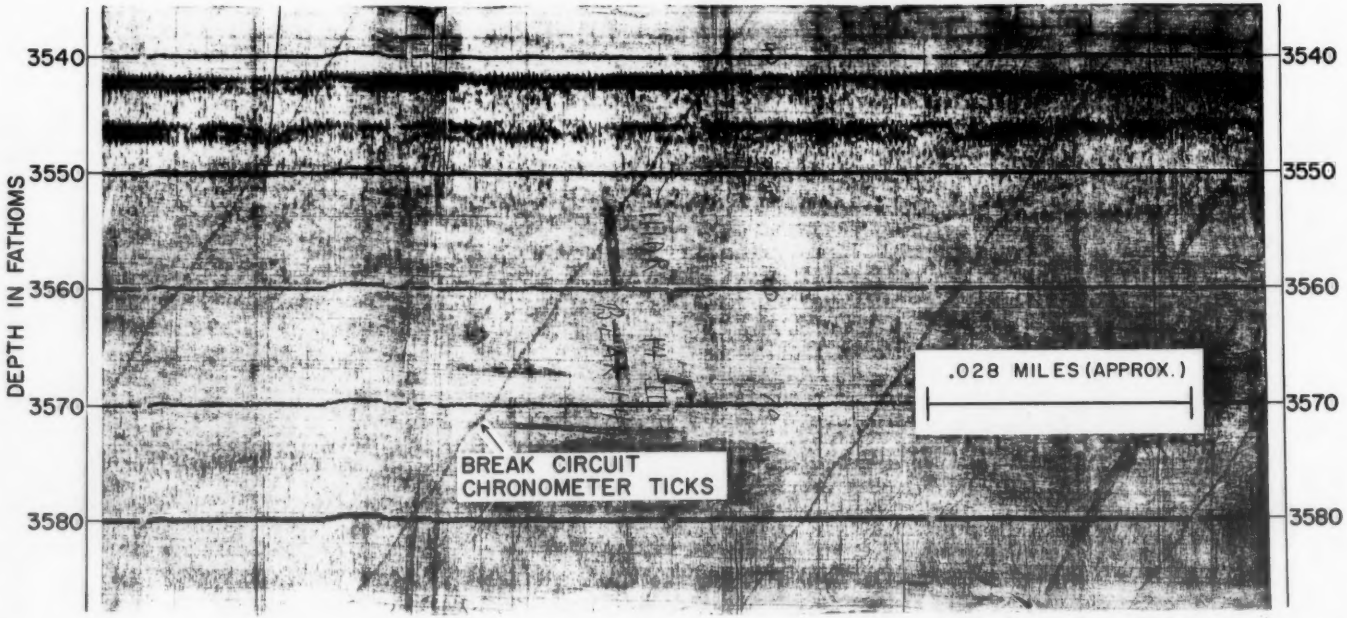
In Plate 6*B* the soundings shown paralleling the edge of the plain are those obtained just within the flat area where traverses crossed its boundary. The soundings shown well out on the plain are either at places where the ship changed course or where a significant change in grade occurred. The sounding lines across the plain show little evidence of undulations even as large as a fathom. However, soundings around the edge show two types of change from rough to flat topography. In one, the depth at the edge is as great as it is farther out on the plain; in the other varying smooth grades are always shoaler at the edge than farther out. Most of these gentle grades appear to be at the mouths of gorges in the surrounding hills; this suggests that these gorges are the supply ducts for sediments that have filled the bottom of the deep. In Plate 6*A* is shown a long sinuous track around the western half of the plain. Here the ship was zig-zagged back and forth across the edge of the flat area in order to cross a number of these shoaler fans. We had no opportunity to explore them in detail, but in the northern central portion a gorge was found trending eastward and gradually shoaling. Farther to the east it deepens irregularly. On the eastern end of the plain, two gorges trend away and shoal gradually to the east. The northern gorge is smooth at its western end where it grades onto the plain, but farther east two sounding lines suggest that its bottom becomes rougher as it shoals. The southerly gorge was surveyed in somewhat more detail (Pl. 6*B*); five sounding

Plate 2.—SLOPE FROM 1000 FATHOMS OFF CUBA TO DEEP PLAIN OF ORIENTE DEEP
Soundings taken from the offing of Santiago de Cuba to the eastern portion of the deep plain

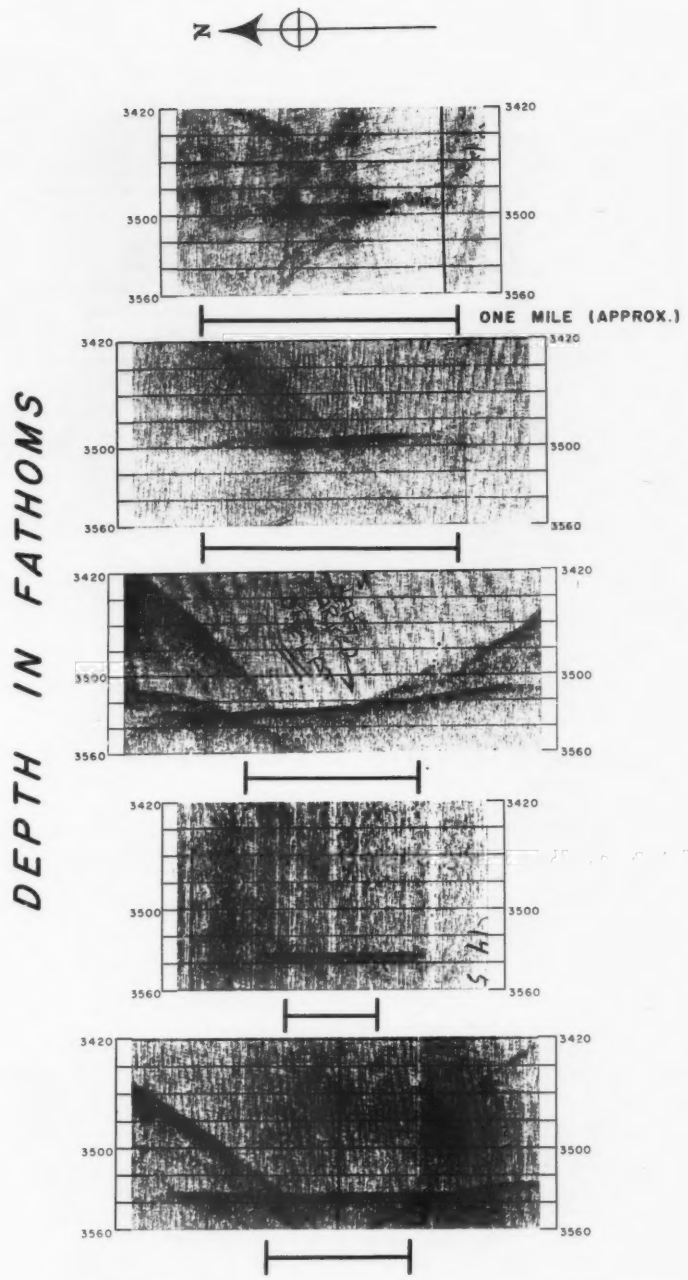
Plate 3.—HIGH-RESOLUTION SOUNDING ON THE FLAT BOTTOM, TAKEN OVER WESTERN PORTION OF DEEP PLAIN
Recordings show multiple echoes probably indicating sound penetration of bottom sediments.



SLOPE FROM 1000 FATHOMS OFF CUBA TO DEEP PLAIN OF ORIENTE DEEP



HIGH-RESOLUTION SOUNDING ON THE FLAT BOTTOM, TAKEN OVER WESTERN PORTION OF DEEP PLAIN



CROSS-SECTIONAL PROFILES OF SOUTHEASTERN GORGE

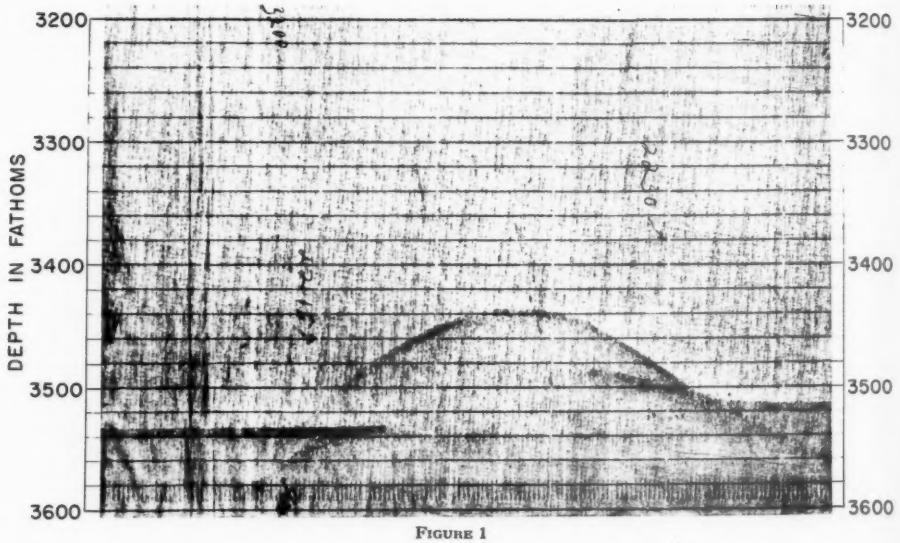


FIGURE 1

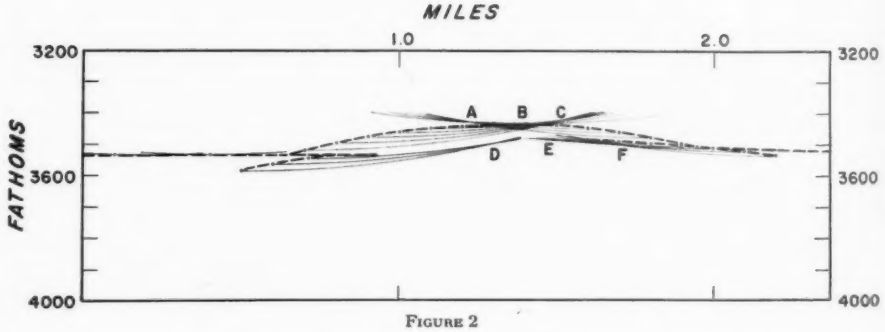


FIGURE 2

DIVIDE BETWEEN NORTHEASTERN AND SOUTHEASTERN GORGES NEAR EAST END OF DEEP PLAIN

Horizontal scales are identical, and figures are aligned.

lines across it are shown in Plate 4. These are arranged from east to west (top to bottom), and the traverses are printed with the northern edge always at the left regardless of the original direction of the traverse. Most of the traverses show pronounced crescent-shaped side-echo sequences from the surrounding rough topography; the smoother bottom of the gorge is shown as a short gently sloping echo sequence. In the upper reaches of the gorge the smooth bottom slopes to the north, but farther west the northerly slope lessens until near its mouth the bottom of the gorge is flat.

Most of the echoes from the smooth bottom of the gorge are somewhat stronger than the side echoes. On one traverse of Plate 4 the noise from the ship recorded by the echo sounder was so great that it obscured nearly all side echoes. Noise interference is a problem in most deep echo sounding; noise intensity depends directly on ship's speed. The worst offender on our ships is the propeller; its rhythmic beat produces the nearly regular pattern of noise evident on all these echograms. This noise never obscured echoes on the flat area but frequently made continuous depth recording underway impossible over the rough topography. This accounts in part for our sketchy knowledge of the topography surrounding the flat. The crescent-shaped echo sequences in the rough topography are almost certainly weaker because the reflected sound is scattered by single highlights rather than reflected from a broad surface such as the flat area.

The echo sequences recorded by an echo sounder do not record faithfully the details of the bottom profile under the ship's track. Echoes may be received from a scattering surface on a rough bottom before and after the ship passes over it, thus producing crescent-shaped echo sequences like those in Plate 5. Such a scattering surface may be laterally displaced from the ship's track, and with present echo-sounding techniques there is no easy way of determining how far it is displaced. Nevertheless if the chip has proceeded on a steady course over such highlights it is possible in some cases to learn more about their shape and size than is obvious from the record. The techniques of this sort of analysis have been discussed by several authors (*e.g.*, Pinke, 1938). Probably the simplest technique is to choose a sounding track perpendicular to the apparent depth contours constructed from the original records. Then on a one-to-one cross section of the traverse, swing a series of circular arcs scaled to the indicated soundings from centers corresponding to the position of the ship for each sounding. The envelope formed by the successive arcs from a given echo sequence may be taken as the approximate profile of the corresponding reflecting surface. Such a profile is not necessarily in the vertical plane under the ship's track, and there may be other complications over complex topography that make the interpretation uncertain. Nevertheless the method gives a more nearly accurate profile than the graphic record and should be applied wherever more clearly defined bottom shapes are needed. As an example, the profile in Figure 1 of Plate 5 was taken across the divide between the two easternmost gorges feeding the flat area. The difference in level of the two smooth portions of the profile indicates that the gorges are separated by the rougher topography. In Figure 2 of Plate 5 the results of an arc-swing analysis are shown. The dashed lines show the recorded soundings scaled to the same vertical scale as the horizontal scale of Figure 1 of Plate 5. The envelopes of the arcs form the closest approximation to actual bottom profiles available from the data. The principal crescent translates to an envelope (*A-B-C*) slightly concave upward, whereas the others form a highlight at *D* and a sloping "surface" *E-F*. The flat profiles, of course, translate to flat profiles. This sample illustrates how different from the form of the sounding record the profile may be and suggests that the rough terrain is characterized by abrupt changes in level, which further suggest that the rough terrain is formed of complexly fractured rock. This interpretation appears the most reasonable of all possibilities because of the generally rugged character of the deep.

PLATE 4.—CROSS-SECTIONAL PROFILES OF SOUTHEASTERN GORGE

Sounding profiles were taken at different speeds over the bottom as indicated by distance scales.

PLATE 5.—DIVIDE BETWEEN NORTHEASTERN AND SOUTHEASTERN GORGES NEAR EAST END OF DEEP PLAIN

Horizontal scales are identical, and figures are aligned.

Figure 1.—Uncorrected profile

Figure 2.—Profile corrected by arc-swinging analysis and with vertical exaggeration removed

Geological Interpretation

The structure of the Bartlett Trough has been inferred largely from its bathymetry and what is known of the structure of Cuba, Jamaica, Hispaniola, and smaller islands in its vicinity. The dimensions of the trough are so great compared with these adjacent islands, and the topography of the trough seems to follow such a unified pattern, that its tectonic history must be related to that of the Caribbean as a whole; interpretations of local structure must be consistent with this fact. Hess (1938) proposed that the trough was part of a more extensive region of differential horizontal movement in which the central Caribbean moves eastward relative to Cuba. Meyerhoff (1954) discussed Caribbean structures in relation to a tectonic hypothesis in which major horizontal couples are proposed as the principal forces acting on the region. Butterlin (1956) reviewed the several principal hypotheses about the tectonics of the Caribbean region and pointed out the differences of opinion about, for example, Cenozoic tectonics in Oriente Province, adjacent to the trough. Students of this region do not agree about structures associated with the southern coast line of Oriente Province; some follow Taber (1931) in regarding the post-early Miocene history as one of subsidence of the trough accompanied by normal faulting; others (Woodring, 1928; Lewis and Straczek, 1955) favor folding and high-angle thrust faulting. All agree that the Bartlett Trough was formed essentially as today by mid-Miocene time and that changes in elevation of the whole region accompanied by faulting have continued to the present.

When combined with other geological and geophysical data about the trough the present study should help to evaluate the several points of view. The topography of the island slope was not studied in detail, but the crossings made indicate interruptions in the slope (Pl. 2). This sounding line was taken about 45° to the general slope. The 16-degree slope along the track from 1000 to 2000 fathoms may be taken as a slope of about 31°, a rather high submarine slope to persist for a mile in depth. The crescent-form echo sequences at the break in slope indicate at least a local reversal in the slope. If a fault zone is associated with this slope, it could be the principal fault zone of vertical movement between the Sierra Maestra and Oriente Deep. A similar fault zone that continues to the second clear reversal at 2800 fathoms might be identified from the beginning of the next steep slope at 2300 fathoms. H. A. Meyerhoff (Personal communication) observed Miocene sediments dipping steeply southward at the coast west of Santiago, which suggests that the slopes just discussed may be bedding-plane slopes. This interpretation further suggests that the Oriente Deep is part of a geosynclinorium, rather than a trough bounded by normal faulting. Pronounced crescent-form echo sequences can be seen from 1000 to 2000 fathoms superimposed on more diffuse sequences (Pl. 2). The diffuse echo sequence indicates fine-scale roughness, whereas the crescent sequences probably come from marked prominences. Sounding lines parallel to the coast show sudden irregular changes in depth. Hence, the slope is probably deeply cut and may be locally talus-covered. Meyerhoff (Personal communication) found transverse faults which "... cut out the Miocene in places, bringing the more complex structures of the Sierra Maestra to the water's edge. Almost invariably these transverse faults have been excavated into deep valleys, and drowned water gaps cut the Mid-Tertiary hogback at subregular intervals." Our data are not adequate to correlate submarine topography with these deep valleys, but such correlation may well obtain. The territory between the slope reversal at 2800 fathoms and the southern limits of the area surveyed has irregular topography and considerable relief (*e.g.*, 400 fathoms in 2 miles within 3–4 miles of the plain), which deepens toward the plain from all sides. We believe that this region is a zone of rock complexly fractured by the same tectonic activity that caused the major faulting and subsidence. The irregular topography of the deep appears to be mountainous rather than smoothly undulating. More bathymetric detail will probably show that this rugged area extends great distances along the Bartlett Trough.

We have suggested that the deep plain is sedimentary, and that the sediment was supplied possibly by density currents flowing down the gorges on all sides. B. C. Heezen (Personal communication) recently took a core 5.5 m long in a similar deep plain south of Guantanamo Bay. His findings strengthen this hypothesis: alternate layers of sand and red clay each about 10 cm thick were found; the topmost layer was sand, and the sand became progressively coarser deeper in the core; at 3 m shell fragments and beach pebbles were found, and, at 5 m, wood. The grades in the gorges are enough greater than those on the plain to suggest that

sediments there are somewhat coarser than on the plain; in one gorge at least (the northern one at the eastern end) rather coarse material or rougher terrain is indicated by the echograms. These gorges appear similar to those bordering the Cul-de-Sac area in southeastern Hispaniola, as described by R. R. Shrock (Personal communication). There a broad valley, partly below sea level, has been uplifted after submergence below the sea (for a more detailed description *see* Butterlin, 1956). The valley floor is flat and covered by a fine sedimentary blanket, in places a fine calcareous ooze. The bordering gorges are cut in limestone cliffs and have smooth bottoms which grade down to the valley floor and are gravel-covered. The qualitative similarity of these two areas suggests that the materials forming them may be similar. For example, the sediment indicated by the multiple bottom echo discussed above may be similar to the calcareous ooze in the vicinity of salt-water lakes in the Cul-de-Sac. However, we know little about the immediate local area other than the facts developed above. A thorough seismic study of the Cayman sea area is needed to reveal the broad structural relations; underwater photography, bottom coring (especially in the deep plain), and continued sounding studies over a much broader area are needed to extend our knowledge of the recent history of the trough.

Although the results presented here are not enough to give full descriptions of the structures, it is reasonable to speculate about the sedimentation taking place in the modern Oriente Deep. The large flat area outlined in Plate 6*B* and the smaller one imperfectly outlined to the west of it appear to be recent sediment having its source in landslides initiated by local seismic activity and probably carried to the greatest available depths by turbidity currents that flow down gorges such as those charted. This type of activity probably has been going on since the subsidence of the Bartlett Trough began. The sediments appear to be ponding in places determined by preexisting topography. The gradual downward trend of rough terrain from all sides into the deep suggests this, but there are probably continuing rock movements in the trough which modify its development in detail. Although no systematic seismic studies of this region have been completed a very few seismic-reflection observations have been made that show a prominent M-type reflection of the sort described by Hersey and Ewing (1949). If this reflection comes from the first solid rock beneath these sediments, its delay of about 0.5 second after the bottom reflection implies a sediment thickness of the order of at least 1200–1500 feet, depending on the compressional wave velocity in the sediments. Test of this deduction will have to await a more systematic seismic study of the area.

Summary and Conclusions

(1) The deep submarine slope off the southern coast of Cuba south of Sierra Maestra is about 31°; it is deeply cut and probably has local talus slopes.

(2) The two principal reversals in slope are at about 2000 fathoms and 2800 fathoms. These may be principal fault zones. Their east-west extent may be only a few miles.

(3) Below 2800 fathoms rugged topography dominates an area about 15 miles wide north-south, and local relief is several hundred fathoms.

(4) The rugged topography deepens gradually to the center of Oriente Deep where there is a nearly flat plain 3–5 miles wide and roughly 25 miles long.

(5) The flat area is bounded by hills and gorges. The gorges have relatively smooth bottoms in their lower reaches which grade onto the flat. These relationships suggest that the plain is underlain by sediment supplied from the surrounding gorges. A core taken recently in a near-by flat area by B. C. Heezen on R/V Vema of Columbia University contains graded sand and clay.

(6) In one part of the plain the character of the multiple bottom echo suggests strongly that secondary echoes come from a subsurface horizon about 30 feet below. Such bottom penetration of 12 kc sound further suggests that the bottom material is a very porous unconsolidated sediment.

(7) Seismic-reflection observations over the plain suggest that the underlying sediment may be about 1200–1500 feet thick.

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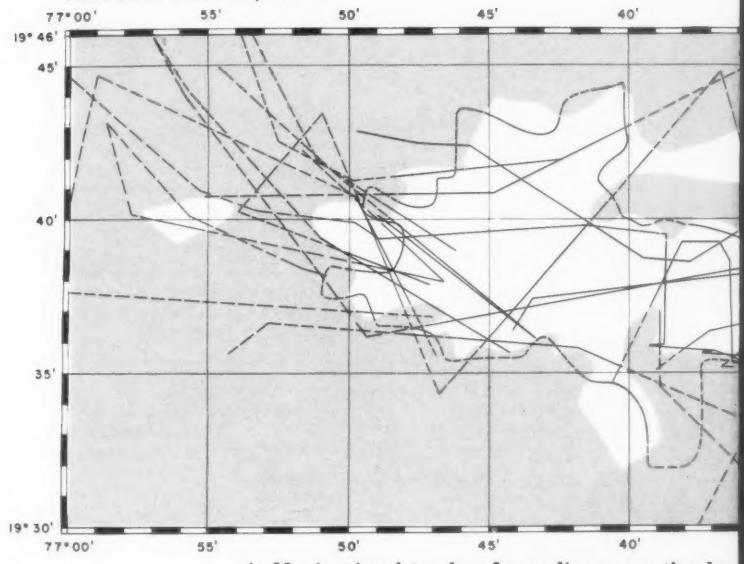
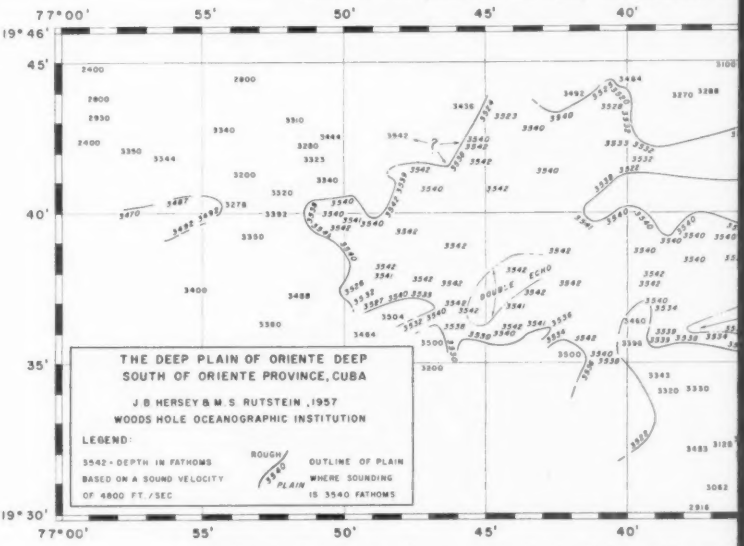
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, DECEMBER 23, 1957

WOODS HOLE OCEANOGRAPHIC INSTITUTION CONTRIBUTION NO. 944



A. Navigational tracks of soundings over the dee
Solid tracks (smooth bottom) and dashed tr



B. Bathymetric chart of the deep plain of the Ori
fee[t] per second sound velocity. Deep-plain so
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shown as a solid line (good control) or a da
sub-bottom echoes were recorded are surrou

NAVIGATIONAL TRACKS OF SOUN
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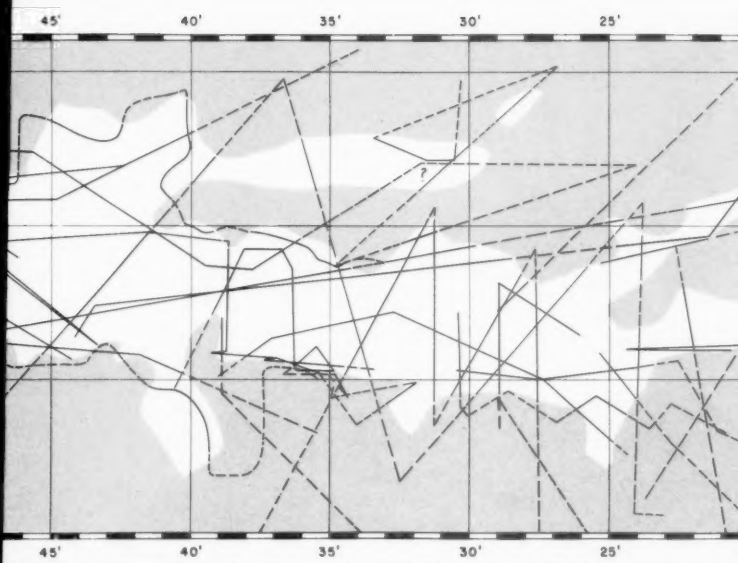
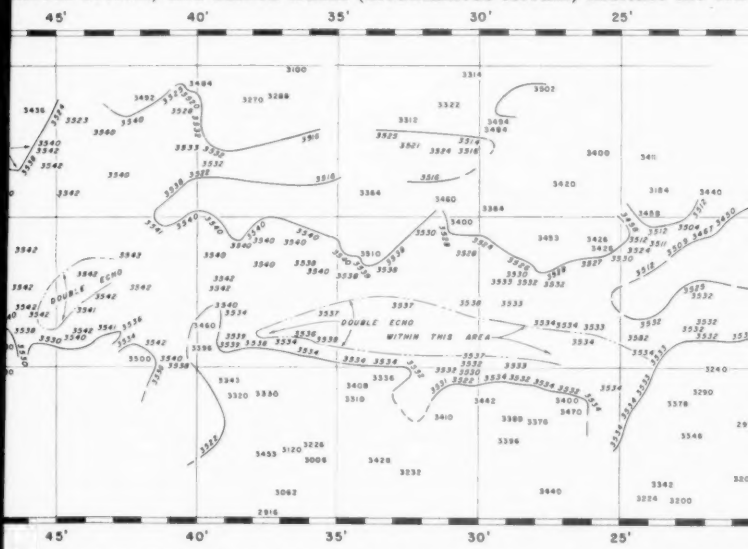
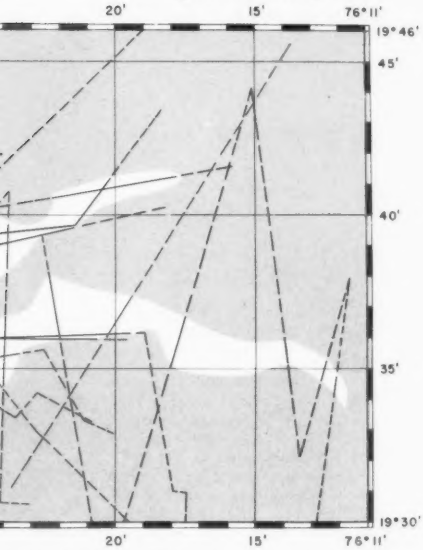


cks of soundings over the deep plain of Oriente Deep. White area is the d
mooth bottom) and dashed tracks (mountainous terrain) indicate the con



rt of the deep plain of the Oriente Deep. Soundings are in fathoms based
sound velocity. Deep-plain soundings (slanted) are separated from selecte
over surrounding mountainous terrain by the observed edge of the de
id line (good control) or a dashed line (poor control). Areas of deep pla
hoes were recorded are surrounded by coded dashed lines.

TRACKS OF SOUNDINGS AND BATHYMETRIC DEEP PLAIN OF ORIENTE DEEP



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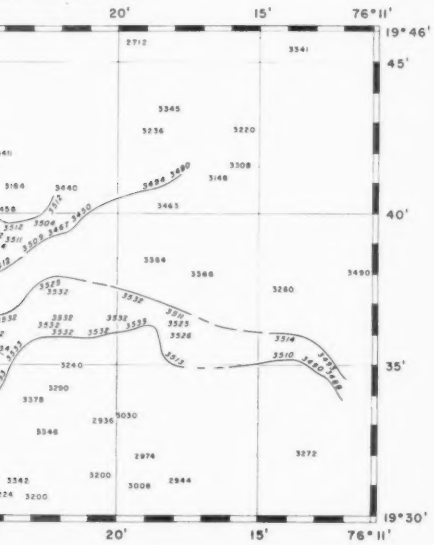

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METRIC CHART,

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			LESS THAN ... DOLOMITE	
QUIET WATER DEPOSITS	FINE-GRAINED LIMESTONE (Calcilutite) Fine-grained limestone with less than 10% sand size or coarser than sand size clastic carbonate fragments *	Carbonate Facies	Fine-grained limestone (1) (including chal...	
		Impure Facies <50% clay-silt size noncarbonate particles	Well indurated	Imp... gra... lim...
			Poorly indurated	Mar...
		Sandy (Arenaceous) Facies <50% quartz sand	Sandy fine-grai... limestone	
	CALCARENITIC LIMESTONE More than 10% sand size or coarser than sand size clastic carbonate fragments * in more than 10% original fine-grained matrix	Carbonate Facies	Calcarenitic limestone (2)	
		Impure Facies <50% clay-silt size noncarbonate particles	Impure calcaren... limestone	
		Sandy (Arenaceous) Facies <50% quartz sand	Sandy calcareni... limestone	
CURRENT WASHED DEPOSITS	CALCARENITE Sand size clastic carbonate fragments * in less than 10% original fine-grained matrix	Carbonate Facies	Calcarenite (3)	
		Sandy (Arenaceous) Facies <50% quartz sand	Sandy calcarenite	
	COARSE CARBONATE CLASTIC (Calcirudite) Coarser than sand size clastic carbonate fragments* in less than 10% original fine-grained matrix.	Carbonate Facies	Coarse carbona... clastic (5)	
		Sandy (Arenaceous) Facies <50% quartz sand	Sandy coarse carbonate clast...	

* Clastic carbonate fragments include sand size and coarser th... size grains irrespective of origin, for example, particles of olde... carbonate rock, ooliths, pellets and biogenic trash.

ARABIAN CARBONATE ROCK CLASSIFICATIO

ORIGINAL TEXTURE NOT VISIBLY ALTERED (except by cementation)		ORIGINAL TEXTURE		
		MODERATELY		
LESS THAN 10% DOLOMITE		LESS THAN 10% DOLOMITE	MORE THAN 10% PORPHYROBLASTIC DOLOMITE	LESS THAN 10 DOLOMITE
Fine-grained limestone (1) (including chalk)		Partially recrystallized fine-grained limestone (11R)	Partially dolomitized fine-grained limestone (11D)	Crystalline limest (12R)
				+Fe: Calcite-Siderite
Well indurated	Impure fine grained limestone	Impure, partially recrystallized fine-grained limestone	Impure, partially dolomitized fine-grained limestone	Impure crystalline limestone
Poorly indurated	Marl	Partially recrystallized marl	Partially dolomitized marl	Crystalline marl
Sandy fine-grained limestone		Sandy, partially recrystallized fine-grained limestone	Sandy, partially dolomitized fine-grained limestone	Sandy crystalline limestone
Calcarenitic limestone (2)		Partially recrystallized calcarenitic limestone (21R)	Partially dolomitized calcarenitic limestone (21D)	Crystalline calcarenitic limes (22R)
Impure calcarenitic limestone		Impure, partially recrystallized calcarenitic limestone	Impure, partially dolomitized calcarenitic limestone	Impure, crystallin calcarenitic limestone
Sandy calcarenitic limestone		Sandy, partially recrystallized calcarenitic limestone	Sandy, partially dolomitized calcarenitic limestone	Sandy, crystalline calcarenitic limestone
Calcarenite (3)		Partially recrystallized calcarenite (31R)	Partially dolomitized calcarenite (31D)	Strongly recrystallized calcarenite (32R)
Sandy calcarenite		Sandy, partially recrystallized calcarenite	Sandy, partially dolomitized calcarenite	Sandy, strongly recrystallized calcarenite
Coarse carbonate clastic (5)		Partially recrystallized coarse carbonate clastic (51R)	Partially dolomitized coarse carbonate clastic (51D)	Strongly recrystallized coa carbonate clastic
Sandy coarse carbonate clastic		Sandy, partially recrystallized coarse carbonate clastic	Sandy, partially dolomitized coarse carbonate clastic	Sandy, strongly recrystallized coa carbonate clastic

and coarser than sand
rticles of older
h.

** Dolomitic and calcitic portions of rock
intimately intergrown as to be generally v
indistinguishable.

ROCK CLASSIFICATION (General)

Bramkamp and Powers, Table 1

ORIGINAL TEXTURE ALTERED				ORIGINAL TEXTURE OBLITERATED
TELY	STRONGLY			
MORE THAN 10% ORPHYROBLASTIC DOLOMITE	LESS THAN 10% DOLOMITE	10-75% GRANOBLASTIC DOLOMITE**	MORE THAN 75% DOLOMITE WITH RELIC TEXTURE	MORE THAN 75% DOLOMITE
ially dolomitized -grained limestone)	Crystalline limestone (12R)	Dolomitic Limestone (12D)	Calcitic Dolomite (13)	CRYSTALLINE DOLOMITE (4) (Impure and sandy facies termed respectively impure crystalline dolomite and sandy crystalline dolomite.)
	+Fe Calcite-Siderite	Calcite - Siderite - Dolomite		
re, partially mitized fine- ned limestone	Impure crystalline limestone	Impure dolomitic limestone	Impure calcitic dolomite	
ially mitized	Crystalline marl	Dolomitic marl		
y, partially mitized fine- ned limestone	Sandy crystalline limestone	Sandy dolomitic limestone	Sandy calcitic dolomite	
ially dolomitized arenitic limestone)	Crystalline calcarenitic limestone (22R)	Dolomitic calcarenitic limestone (22D)	Calcarenitic dolomite (23)	
re, partially mitized arenitic limestone	Impure, crystalline calcarenitic limestone	Impure, dolomitic calcarenitic limestone	Impure calcarenitic dolomite	
y, partially mitized arenitic limestone	Sandy, crystalline calcarenitic limestone	Sandy, dolomitic calcarenitic limestone	Sandy calcarenitic dolomite	
ially mitized arenite (31D)	Strongly recrystallized calcarenite (32R)	Dolomitic calcarenite (32D)	Calcarenite dolomite (33)	
y, partially mitized arenite	Sandy, strongly recrystallized calcarenite	Sandy dolomitic calcarenite	Sandy calcarenite dolomite	
ially dolomitized se carbonate tic (51D)	Strongly recrystallized coarse carbonate clastic (52R)	Dolomitic coarse carbonate clastic (52D)	Coarse carbonate clastic dolomite (53)	
y, partially mitized coarse onate clastic	Sandy, strongly recrystallized coarse carbonate clastic	Sandy, dolomitic coarse carbonate clastic	Sandy coarse carbonate clastic dolomite	

Dolomitic and calcitic portions of rock equigranular and so mately intergrown as to be generally visually stinguishable.

(2) Indicates equivalent reservoir rock type

Short Notes

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CLASSIFICATION OF ARABIAN CARBONATE ROCKS

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INTRODUCTION

The classification of carbonate rocks used by the Arabian American Oil Company was developed for the study of carbonate rocks in Arabia. The results, primarily covering shallow-water carbonates, seem to have relatively general application. The basic material studied has been about 25,000 feet of cores, 30,000 core plugs, extensive Jurassic outcrops in Central Arabia, and considerable work on modern Persian Gulf sediments during the last twenty years.

Adequate classifications to serve the various purposes required by the Arabian American Oil Company have not been found in the literature. As a result, it has been necessary to develop a classification which not only presents the nature of the carbonate sediments at the time of deposition, but also goes sufficiently into the changes they have undergone after deposition to allow adequate description for use in reservoir and other detailed geological studies. The classification as presented (Table 1) is a summation of our experience with carbonate rocks, and as such it combines the ideas and opinions of many geologists who have worked in Arabia.

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DISCUSSION

A number of classifications have been proposed for the carbonate rocks. Often, the tendency has been to treat them as a unique suite of rocks somehow divorced from normal sedimentary mechanics. An exhaustive study of the Arabian Jurassic outcrop, which forms a belt of almost continuous exposures nearly 1000 km in length, and 70 or more suites of diamond cores from the Jurassic Arab D carbonate reservoir, as well as preliminary studies of modern calcareous Persian Gulf sediments, has shown that, just as with the siliceous rocks, a natural set of relationships exists between environment and original rock texture. Consequently, the writers feel that first and foremost any classification of carbonates must deal with them as sediments.

Just as in the regular shale-sand-conglomerate suite of rocks, particle size and sorting offer the obvious first attack on the problem of the current regimen under which deposition took place. This would appear to be true in spite of the chemical or biogenic nature of most of the particles. The size and sorting exhibited by the final deposit will in general be equally reliable indicators of depositional environment whether the particles formed in place, near by, or were transported from some distant source. Once this apparently rather formidable barrier in thinking on carbonate rocks has been breached, their systematic classification is a simple matter.

Recent conclusions on sediments of the continental shelf (Dunbar and Rodgers, 1957, p. 48–55) suggest that the problems in siliceous sedimentation are much closer to those of carbonate rocks than previously supposed. To be sure, the range of sediments may differ from siliceous ones, there are no carbonate particles with the flotation properties of layered hydrophyllic clay materials, and, siliceous sediments do not aggregate into organic reefs; if a few qualifications are applied, however, it seems that the overall conclusion is inescapable.

Few carbonate rocks can be disaggregated and sieved as can so many siliceous sediments.

* Deceased, September 1, 1958

It is apparent therefore that in practical application descriptive terms of less precision will have to be used most of the time—reserving very time-consuming quantitative thin-section studies for research projects. In any case there is usually much doubt as to which carbonate particles of silt size and smaller have remained unaltered in size and shape since original deposition. For these reasons practical classification needs to be based on visual estimates of the size and sorting of particles, supplemented by more precise studies as these are feasible.

After classification of the rocks according to particle size the next logical subdivision appears to be to break them down further according to the degree to which their original texture has been altered by diagenetic and post-lithification processes. Some carbonates have had the original sedimentary texture obliterated by intense dolomitization. Nevertheless, ultimate gaps in the over-all pattern of sedimentation from causes such as this are usually not large.

Obviously, no classification can be considered entirely satisfactory for in practice complete gradational series exist between end-member rock types. The classification presented here is based on empirical evidence and is known to be functional for work with Arabian carbonates. Preliminary work on outside samples and an extensive survey of literature on carbonate rocks indicate that it may prove to have as wide an application as any scheme yet proposed. Rock types not extensively represented in Arabia such as clastic dolomite sands and organic reef complexes can readily be included with no modification of the basic classification framework.

General Classification

Based on their original particle size, the carbonate rocks may be divided into four main groups: (1) fine-grained limestone; (2) calcarenitic limestone; (3) calcarenite; and (4) coarse carbonate clastic. As the first two types are normally quiet-water deposits and the third and fourth are related to current-washed regiments, they are discussed under these headings. A third environment, the organic reef, is also considered. Any given carbonate rock may be a composite of several of these genetic types. Any of these deposits may undergo limited to intense alteration by recrystallization and dolomitization.

Quiet-Water Deposits

Quiet-water deposits are those clay-silt-size sediments which accumulate in sheltered areas near shore or in depths farther basinward. In connection with this, depth of water does not now appear to be as rigid a controlling factor for particle-size accumulation as once believed. Coarse deposits have been found at abyssal and even bathyal depth, and finer sediments occur near shore. Basic fractionization appears to be brought about by differential current effects.

A large number of carbonate rocks are exceedingly fine-grained, as they are composed of clay-silt-size particles of calcium carbonate. These rocks are the equivalent of siliceous clays and silts except for composition. Because of the limited variation in texture and composition, these rocks cannot be genetically classified with any degree of assurance. It is probable that many of them were deposited as detrital calcite grains in the textural range of silt and clay, whereas others were precipitated as lime muds. When dealing with this textural range it is extremely difficult to determine whether the particles result from abrading and transport or direct chemical or biochemical precipitation. Regardless of origin, however, both are indices of quiet-water deposition. It is well understood that lime muds are at times deposited in association with carbonate fragments considered indicative of current-swept environment; this is particularly true in the case of coarse skeletal material (coquina, calcirudite, etc.). However, it is not at variance with the basic quiet-water premise in that heavy organic growth, including grasses, tend to trap the fine sediment and provide shelter against winnowing current action.

In most examples of fine-grained texture, the very small euhedral calcite crystals cannot be examined individually as they are indistinguishable under the low-power binocular microscope. Quiet-water deposits are divided into two groups on the basis of texture: fine-grained limestone and calcarenitic limestone.

Fine-grained limestone and calcarenitic limestone.—Studies of modern lagoonal sediments indicate that fine-grained limestone and calcarenitic limestone are products of the same or similar depositional environments. The difference, as stated above, is textural.

Fine-grained limestone is essentially equigranular. For the purpose of this classification, the term is used to designate clay- and silt-size carbonate rocks with less than 10 per cent sand-

size or coarser than sand-size clastic carbonate fragments. Because of uniform particle size, grains themselves are the basic textural element.

Calcarenitic limestone is primarily a fine-grained limestone to which more than 10 per cent clastic carbonate fragments has been added. The carbonate fragments may form as much as 90 per cent of the rock; the critical criterion is that 10 or more per cent of the original fine-grained matrix is still intact. In some instances, the carbonate fragments are coarser than sand size. Genetically related to calcarenitic limestone, these rocks are termed coarse carbonate clastic with fine-grained matrix. The textural elements most commonly associated with calcarenitic limistone are grains and matrix and, rarely, in limited quantity, visible cement.

The fine-grained and calcarenitic limestone groups may be separated into three facies on the basis of composition. The criteria for each of these divisions (carbonate, impure, sandy) are as follows:

(1) Carbonate Facies—Components of rock appear to be exclusively carbonate minerals; no impurities recognizable

(2) Impure Facies—Contains from recognizable amounts to 50 per cent clay- and silt-size noncarbonate impurities; carbonate minerals constitute 50 per cent or more of rock

(3) Sandy Facies—Fragments of detrital minerals (dominantly quartz) ranging in abundance from a few grains to 50 per cent; remainder of rock is carbonate minerals

Although uncommon, it is probable that both clay-silt and sand may be admixed simultaneously with carbonate minerals to form an impure and sandy carbonate. As before, the carbonate minerals must be the dominant constituent.

Current-Washed Deposits

Current-washed deposits are those coarser clastic sediments that tend to accumulate in exposed areas near shore, on bottom-high shelf areas, or in deeper water where they are subjected to current action.

Many carbonate rocks are composed almost exclusively of well-washed sand and coarser than sand-size clastic carbonate fragments. Finer particles, responding to current energy, have by-passed or have been removed from this agitated environment and deposited in more sheltered areas. It is this removal or nondeposition of the finer fraction that is the primary means of differentiating between calcarenitic limestone and current-washed carbonate. Except for composition, these rocks are the equivalent of noncarbonate sands which have been deposited on current-washed bottoms. Like other clastic deposits, these lime sands exhibit mechanically produced textures and structures which indicate that they have been transported and more or less abraded and sorted before final deposition. After deposition, most clastic carbonates tend to be poorly to strongly lithified by the chemical precipitation of crystalline calcium carbonate in the pore space. Current-washed deposits are texturally divisible into two main groups: calcarenite and coarse carbonate clastic.

Calcarenite and coarse carbonate classtic.—The term *calcarenite* is used to describe detrital carbonate rocks composed of sand-size (1/16 to 2 mm.) clastic carbonate fragments. For the source of these fragments the reader is referred to the section on Elements of Carbonate Sedimentation. Calcarenite may be with or without authigenic calcite cement. The possible gross textural elements are grains and cement.

When more than half the clastic carbonate fragments of the rock are coarser than sand size, it is called a *coarse carbonate clastic.* With few exceptions, these coarser-textured rocks have significant percentages of interstitial sand-size carbonate fragments. The term *calcirudite* has been proposed for the coarser-grained deposits. Unlike calcarenite, it has failed to gain widespread acceptance. Possible textural elements include grains, cement, and, where sand-size particles form more than 10 per cent of the rock, matrix. Calcarenite and coarse carbonate clastics may be composed predominantly of one type of clastic carbonate fragment, for example oölite, or contain an admixture of several types, such as oöliths, biogenic trash, and fragments of older rock. Where one type predominates, the rock may be so qualified—*i.e.*, oölite calcarenite, biogenic coarse carbonate clastic (coquina), etc. Where the particular organism associated with a biogenic carbonate can be identified, the rock may be more accurately designated—for example, foraminiferal calcarenite, brachiopod coarse carbonate clastic, etc. Each clastic carbonate group has two compositional variants: a carbonate facies and a sandy facies. Significant amounts of clay- and silt-size particles are likely to have been eliminated in the current-swept environment in which these rocks are laid down. Consequently, an impure facies is not considered probable.

(1) Carbonate Facies—Rock is composed exclusively of sand and coarser than sand-size carbonate grains.

(2) Sandy Facies—Detrital noncarbonate sand (from a few grains to 50 per cent) has been admixed with carbonate fragments.

Organic-Reef Carbonates

The organic-reef carbonates, variously called autochthonous, accretionary, and biohermal, are formed by permanently attached, lime-secreting organisms. The general carbonate-rock classification chart does not include organic-reef carbonates primarily because rocks of this type, when recovered as cuttings from widely spaced wells, cannot be distinguished from current-washed deposits. It is only in outcrop, where much of the structure is visible, that the qualifying differences are apparent. Where identifiable, carbonate rocks formed by sedentary reef-building organisms are of sufficient importance to be briefly discussed here.[1]

These organic structures, composed mainly of attached organisms such as corals, brachiopods, algae, stromatoporoids, etc., show considerable variation in size and shape. They aggreagate, along with trapped debris, into moundlike structures measured in inches or thousands of feet.

Fossil-reef environments were undoubtedly similar to that in which their modern counterparts are growing. The reef area is subjected to strong sweeping action by clear, well-aerated water. The organisms, particularly encrusting forms, build a rigid framework or reef core in which broken organic debris and other loose clastic material become lodged and consolidated. During periods of active growth, the reef expands against the current. Whether growth is mainly upward or basinward, or some component of these directions, is dependent in large part on the relative stability of sea level.

Commonly, carbonate fragments, in large part derived from the reef core, accumulate outward in the form of a clastic fan. Many of these deposits show evidence of some transportation, sorting, and stratification and thus are genetically related to the current-washed deposits. Recent findings indicate that calcareous algae, and not coral as commonly supposed, are often the dominant form in the organic reef; however, this may not apply in all coral-reef areas.

[1] For a detailed analysis of the organic reef rocks, the reader is referred to N. D. Newell *et al.* (1953).

Carbonate-Rock Sedimentary Texture

Any of the quiet-water, current-washed, and organic-reef carbonates may remain essentially unaltered or may undergo diagenetic modification and consequent changes in primary sedimentary texture and composition. The original calcite is commonly replaced by dolomite and in some instances by siderite. All available evidence indicates that the dolomite associated with normal Arabian carbonate rocks is the result of such replacement rather than of primary precipitation. However, this does not take into account the possible presence of an unknown, perhaps considerabe, amount of primary dolomite in strongly evaporitic facies.

Recrystallization may alter texture and composition (dolomitization) or merely alter texture without changing the composition. Such changes are primarily a result of reactions between original sediments and contained fluids. The processes by which calcite sediments are made dolomitic are discussed in the section covering post-lithification changes. Based on the lack or intensity of alteration of the original sedimentary texture, the carbonate rocks are divided into four types: (1) unaltered sedimentary texture, (2) moderately altered sedimentary texture, (3) strongly altered sedimentary texture, and (4) obliterated sedimentary texture.

Unaltered sedimentary texture.—The original sedimentary texture of rocks of this type has not been visibly altered except for compaction and cementation. The rock is considered essentially unaltered if dolomite does not exceed 10 per cent. The compositional and textural make-up of these rocks, as well as the terminology used, is the same as that just described under noncurrent and current-washed deposits. These include fine-grained and calcarenitic limestone and calcarenite and coarse carbonate clastic. The one exception is that the impure facies of the fine-grained limestone has been separated into two subtypes on the basis of induration. These are marl and impure fine-grained limestone. As here defined, marl is a poorly indurated fine-grained carbonate rock which contains up to 50 per cent clay- to silt-size impurities. When wet, marl generally becomes plastic or claylike.

One common texture associated with the unaltered carbonates results when crystal faces

are not fully developed during diagenesis of lime muds or crystallization from a fluid. These incomplete crystals are easily recognized, and the texture developed is called *crystalline granular*.

Moderately altered sedimentary texture.—Where original carbonate sedimentary texture is essentially intact but shows evidences of obvious recrystallization, the rock is termed moderately altered. This type of alteration can be brought about by two processes: (1) recrystallization alone without change in composition; or (2) recrystallization involving dolomitization. The first process is called partial recrystallization, the second partial dolomitization.

Partially recrystallized texture is brought about by the slight alteration of original components to a poorly developed or patchy calcite mosaic. Little or no dolomite is added (less than 10 per cent). The limited degree of alteration does not interfere with the recognition of the original rock texture. This lack of intensity in destructive alteration, coupled with noticeable effects of recrystallization, is used to distinguish partial recrystallization. The effects of this type of alteration have been recognized when superimposed on fine-grained through coarse carbonate textures.

Partially dolomitized texture results from recrystallization due to the addition of more than 10 per cent authigenic calcium magnesium carbonate in the form of discrete euhedral rhombs or porphyroblasts. The individual rhombs are readily visible, generally against a background of slightly altered original texture. Rocks of this type include partially dolomitized fine-grained and calcarenitic limestone and partially dolomitized calcarenite and coarse carbonate clastic.

Moderate post-depositional changes in fine-grained limestone often result in a porphyroblastic texture in which conspicuous medium-grained crystalline to coarsely crystalline dolomite rhombs are scattered throughout the original finer-grained matrix. In some cases the euhedral rhombs are in the same size range as the matrix, and the rock is nearly equigranular. The dolomite rhombs are visible as well-formed crystals in an anhedral crystalline background. These dolomite rhombs may be selective and concentrate in patches or be uniformly scattered throughout the rock.

Partially dolomitized calcarenitic limestone may show very selective growth patterns. In some, rhombs are confined to the fine-grained matrix; in others, both matrix and clastic carbonate grains have been affected. The obvious secondary nature of the dolomite is often displayed by rhombs cutting across the boundary between grains and matrix.

When calcarenite and coarse carbonate clastic rocks have undergone partial dolomitization, the dolomite is distinguished from the original rock and secondary cement by its generally darker color and discrete rhombohedral form. The dolomite may be scattered throughout, replacing or filling voids left in clear calcite cement. It may cut across the boundary of the cement and grains or, in some, be confined to or within the grains. The latter is particularly evident in coquina in which the internal part of shells has been filled with a mosaic of individual dolomite rhombs and calcite cement.

Strongly altered sedimentary texture.—A carbonate rock is strongly altered when the original texture, though still recognizable, shows serious destruction by recrystallization to an anhedral calcite mosaic and/or equigranular (granoblastic) dolomite.

On the basis of method and degree of recrystallization three types of strongly altered textures are recognized: (1) crystalline or strongly recrystallized, (2) dolomitic, and (3) relic.

Crystalline or strongly recrystallized texture shows notable changes in original texture. This may be brought about simply by the rearrangement of calcium and carbonate ions through solution and reprecipitation. When this occurs with sufficient intensity to disrupt original textures seriously, but without the addition of significant amounts of dolomite, the rock becomes crystalline or strongly recrystallized. The original texture is replaced by a mosaic of finely to coarsely crystalline calcite. Crystalloblastic textures may arise through the growth of larger calcite crystals at the expense of smaller ones or by conversion of aragonite to calcite. Less than 10 per cent dolomite is involved in the replacement processes in the development of crystalline limestone and calcarenitic limestone and strongly recrystallized calcarenite and coarse carbonate clastic.

Fine-grained limestone is particularly susceptible to recrystallization and is converted to a microcrystalline to coarsely crystalline mosaic of sutured crystals. Where ferrous-rich solutions have been active in the alteration, calcite-siderite rocks develop. Many of the impure and sandy facies show particles of clay-silt-size impurities or quartz grains as inclusions in a glassy mosaic of anhedral crystals.

The fine-grained matrix of calcarenitic limestone is recrystallized in the manner described above. The texture of the clastic carbonate fragments may or may not be altered. In some cases only the internal structure has been recrystallized, whereas in others recrystallization of the inclusions is complete, and crystalline cleavage extends uninterruptedly across matrix and grains.

Three principal variations in texture have been noted associated with strongly recrystallized calcarenite and coarse carbonate clastic. First, the internal structure of carbonate grains, especially oöliths, has been destroyed by sutured crystals of calcite. The cement and edges of the grains are virtually unaltered. In the second variation the fragments are completely replaced by a mosaic of crystalline calcite not crystallographically related to the cement. In the third, cleavage passes uninterruptedly from replaced carbonate fragments to clear calcite cement.

Owing to extensive recrystallization, it is sometimes impossible to determine whether the rock was originally calcarenite or coarse carbonate clastic.

Dolomitic texture defines rocks which contain 10–75 per cent granoblastic dolomite in which the dolomitic and calcitic portions of the rock are equigranular and generally visually indistinguishable. Our own practical experience with Arabian rocks leads us to prefer these particular limits. The calcitic and dolomitic part of the rock may form a microcrystalline to coarsely crystalline mosaic of anhedral to subhedral crystals. In other rocks, the dolomite may be confined to patches producing a brecciated appearance. Although, in many instances, original textures tend to be obscured by this process, it can generally be recognized in chips without the aid of thin section by examination of the rock as a whole. The equigranular character and inherent difference in degree of alteration seem to distinguish the dolomitic carbonates from those which have been partially dolomitized, although the percentage of dolomite may be the same.

The fine-grained limestones are altered to dolomitic limestone and marl and to calcite-siderite-dolomite rock. The texture is generally a finely to coarsely crystalline granoblastic mosaic of dolomite and calcite. The dolomite may be in patches or uniformly admixed with calcite. The calcite-siderite-dolomite rocks have been altered by the addition of both iron and magnesium. Whether these are added simultaneously is unknown. As in crystalline limestone, it is often possible to detect clay-silt or quartz-sand impurities "floating" in the crystalline groundmass.

Dolomitic calcarenitic limestone rocks are often the most difficult to classify accurately. Visual examination alone will commonly result in their being typed as a dolomitic calcarenite. Where doubt exists, the best approach is to prepare a polished or thin section. If the original rock was calcarenitic, the clastic carbonate grains, although replaced, are generally fairly well preserved and separated by a finely recrystallized matrix. If the rock was originally a calcarenite, the crystalline "matrix" is discontinuous because of the proximity of the grains. Some calcarenitic limestones exhibit a pattern of selective dolomitization, and dolomite is restricted to grains or to matrix. In others the alteration is indiscriminate.

The dolomitization of calcarenite is among the most interesting and complex of any of the carbonates. The internal parts of shells and oölites may be filled with a crystalline mosaic of dolomite enclosed by crystalline calcite. In others the enclosed dolomite is intimately intergrown with crystalline calcite. In still others dolomite replacement is restricted to specific types of carbonate fragments.

Relic texture is seen where carbonates have been subjected to intense alteration through the addition of 75 per cent or more of dolomite but vestiges of original texture sufficient for identification are retained. In some cases the original textures are fairly well preserved by replacement, but generally only vague, obscure remnants are still visible. It is this retention of original texture which differentiates these rocks from crystalline dolomite in which the original texture is unidentifiable. Incomplete conversion of the fine-grained rocks to dolomite produces a porphyroblastic texture. The grains are medium-grained-crystalline to coarsely crystalline, rhombic euhedra of dolomite set in 25 per cent or less of fine-grained or finely crystalline calcite matrix. The fine-grained matrix limits the porosity of the rock to essentially that of unaltered fine-grained limestone. In some calcitic dolomites, the dolomitization is patchy, and small areas of unaltered fine-grained limestone remain. The impure and sandy facies is often detectable where dolomite rhombs enclose clay-silt and sand admixtures.

The description of dolomitic calcarenite is valid for calcarenitic dolomite except that modification is more advanced with a subse-

quent change in percentage of dolomite (more than 75 per cent). It is often necessary to use polished or thin sections on specimens of this and other strongly altered types to calibrate the visual examination of rock chips. Once this has been done it is not too difficult to detect the "ghosts" of clastic carbonate fragments imbedded in a finely crystalline matrix.

A few specimens of calcarenite dolomite have the original texture preserved clearly even though more than 75 per cent of the original rock has been replaced by dolomite. For the most part, however, the clastic texture is so completely altered that it is visible only after close examination of the entire specimen. A hint of the original clastic texture may come from small, less altered patches. In rocks of this type all clastic carbonate fragments may be replaced, with only a trace of original clastic texture remaining, but good evidences of clear calcite cement are retained. In others, the calcite cement and most of the clastic carbonate fragments have been replaced, leaving only intergrowths of calcite crystals.

Obliterated sedimentary texture.—A rock shows obliterated sedimentary texture when dolomite exceeds 75 per cent and original texture can no longer be determined with any assurance. Some rocks of this type still retain within the dolomite crystal network what appears to have been an original fine-grained constituent; however, it is not possible to determine whether this represents original matrix or intensely recrystallized cement and/or clastic fragments of a calcarenite or coarse carbonate clastic.

Crystalline dolomite is commonly composed of medium-grained-crystalline to coarsely crystalline anhedra exhibiting a granoblastic texture. A less common texture is porphyroblastic in which distinctive medium to coarse crystalline rhombs are imbedded in a finely crystalline dolomite matrix. Where impure and sandy carbonates have been completely altered, the rhombs enclose the impurities.

Elements of Carbonate Sedimentation

Needless to say, a rigorous analysis of the processes and environmental factors associated with carbonate deposits would be a major undertaking beyond the scope of this work. Consequently, only a brief outline along with some of the more fundamental aspects will be considered. The reader is referred to other sources for a more detailed treatment of this subject (Krumbein and Sloss, 1953).

The characteristics peculiar to a carbonate deposit are determined by the complete environmental framework within which the particles have been derived, transported, deposited, and lithified.

Source

As stated previously, the source of the fine-grained carbonate components which make up the lime muds is usually uncertain. Although some carbonates may have been chemically or biochemically precipitated in place, many are considered to have been formed by the abrasion of clastic carbonate fragments to clay and silt size. If these clastic fragments are sand or larger than sand size, their origin can often be determined. They are derived from four sources: (1) older carbonate rocks, (2) biogenic or organic material, (3) oöliths or pisolites, and (4) pellets.

(1) *Older carbonate rocks* include any carbonate rock that has previously undergone lithification. Subsequently, the rock is broken down, transported, and deposited, often with much of its original character intact. In some Arabian rocks, fragments of older calcarenite may be enclosed as part of a later calcarenite or coarse carbonate clastic. Although of limited distribution in the Arabian sequence clastic dolomite sands are common elsewhere and are basically only a particular type of calcarenite derived from pre-existent rock.

(2) *Biogenic sediments* are composed of organic remains regardless of type. Examples include organic reefs and associated reef talus, shells and shell fragments, fossil oozes, chalk, etc.

(3) *Oöliths* are rounded (spheroidal or ellipsoidal) sand-size carbonate bodies which in cross section show concentric structure. If the carbonate body is larger than sand size (2 mm), the term *pisolite* is used. Oöliths appear to have originated in shallow, strongly agitated water.

(4) *Pellets* are rounded (spheroidal or ellipsoidal) sand-size forms with no internal structure. This lack of structure distinguishes the pellet from the oölith. The composition is generally a heterogeneous aggregate of clay- and silt-size fragments of carbonate and quartz. The origin of these pellets is uncertain, but one assumption is that some represent fossilized faecal pellets.

The noncarbonate elements are primarily derived from the weathering (chemical or physical destruction) and transport of a terrigenous rock. These impurities are generally

quartz in the sand- and silt-size range and clay minerals in the smaller particles.

Transportation

Essentially, in Arabian deposits the transportation of the carbonate minerals appears to have been limited to wave and current action. The minerals are moved by these agencies in solution, in suspension, by saltation, and/or by rolling. Limited chemical alteration may occur during movement; however, more important, repeated collisions of the individual grains with other particles and the bottom cause change in size, shape, and roundness. These characteristics are reflected in the final deposit as sediment sorting is related directly to particle size, shape, and relative density. Because the effects of abrasion and sorting are the ones preserved, they are the only one mentioned here. This is obviously a simplified treatment of a complex subject in which factors such as dynamics of fluid motion, propulsion of granular materials, settling velocity of particles, etc., are all important.

The clay-silt and sand impurities were probably carried by running water and wind to the edge of the depositional basin where wave and current action took over.

Deposition

Stated briefly, a particle is deposited when fluid energy can no longer displace or move the particle. Basically, however, this is but a small part of the over-all depositional picture. When viewed regionally it is seen that facies or type of sedimentation is controlled in turn by sediment supply, depositional environment, basin configuration, and over-all basinal tectonic framework.

Most investigators choose to divide sedimentary environments into marine, continental, and a mixed or transition environment. The marine environment is divided on the basis of depth into neritic, bathyal, and abyssal. As most carbonate rocks of Arabia were probably deposited in or near the neritic zone, this is the general environment with which we are most concerned.

Although no fixed rules pertain to the deposits of the neritic and near-neritic environment, it is apparent that current-washed deposits (calcarenite and coarse carbonate clastic) will often, though by no means always, be the sediment of the relatively shallower parts. The agitated parts of the environment include near-shore and bottom-high or shelf areas farther basinward as well as deeper parts of the basin subjected to scouring currents. The fine-grained and calcarenitic limestones will be more or less restricted to the deeper or more sheltered areas. The organic-reef limestones are products of the current-swept environment.

The facies picture is complicated by the addition of clay-silt and sand impurities into the depositional basin. Primarily, sand impurities are concentrated near shore, and the clay-silt fraction is carried farther out to be dropped with the quiet-water deposits.

Diagenesis and Lithification

Diagenesis.—Diagenesis is generally considered to include those physico-chemical changes which take place in a sediment after deposition and prior to lithification. These alterations in large part appear to have been brought about by the marked changes which occur in the chemistry of connate solutions when they are no longer able to circulate freely with the overlying sea water. After entrapment by consolidating sediment, sea water is subject to disequilibria-producing shifts in oxidation-reduction potential, hydrogen-ion concentration, amount of dissolved gases and salts, bacterial content, etc.

Lithification.—Lithification is defined as those processes or combination of processes which consolidate and indurate a sedimentary deposit. It is apparent that changes originating during diagenesis are involved in rock induration and continue to be active after lithification. For this reason, the processes associated with diagenesis and lithification are not here considered separately. Although diagenetic and post-lithification changes which take place in carbonate sediments are numerous, the more important modifications appear to be compaction, cementation, recrystallization, dolomitization, and authigenic addition of minerals.

Compaction results when rock particles are pressed together by the weight of overlying sediment. The critical alterations are elimination and forced migration of water and reduction of pore space. Fine-grained limestone is an excellent example of a rock in which little alteration other than compaction has taken place.

Cementation of carbonate rocks results most commonly by the chemical precipitation of crystalline calcite around grains and matrix.

Other less common cementing agents include dolomite, siderite, silica, and, it is believed in some instances, anhydrite. The cement may be derived from an outside source, from connate water, and from solution between grains at point of contact and redeposition in adjacent pore space. Again the primary change due to cementation is decrease in effective porosity.

Recrystallization is the textural alteration, with or without accompanying change in composition, of previously lithified rock generally brought about by the increase in size of component crystals.

Studies indicate that the process involves the growth of larger crystals at the expense of smaller ones. Carbonate studies in Arabia, coupled with a survey of the standard reference works, show that recrystallization most often is accomplished in two ways. The first involves only a rearrangement of existing calcium carbonate, the second dolomitization. Known evidence for primary dolomite is lacking in Arabian rocks. On the contrary, there are considerable data to indicate that original calcite has been made dolomitic by replacement. This is particularly evident in carbonates with a small to dominant amount of sand-size particles. In these, dolomite often is superimposed and transects original texture. Dolomitization alters texture and composition as well as such basic reservoir properties as porosity and permeability.

Authigenesis involves *in situ* mineralization at the time of, or after, the deposition of a sediment. This includes the crystalline calcite cement which is so common in the coarser carbonates. Dolomite associated with Arabian carbonates appears to have been added exclusively in this manner. Numerous other constituents are considered authigenic such as glauconite, euhedral quartz, siderite, and pyrite.

Classification of the Reservoir Rocks

To function as a reservoir, a rock must have void space in which hydrocarbons can accumulate. To serve as an economically adequate reservoir rock, the pore space must be sufficient to contain commercial quantities of oil and be interconnected or permeable enough to give up the oil with relative ease.

The basic framework of the general carbonate rock classification can readily be applied with but slight modification to the classification of the Arabian carbonate reservoir rocks (Table 2). The basic change between the two classifications is the expansion of the clean carbonate facies into three subtypes based on visual porosity differences. Observed reservoir rocks have contained no recognizable amounts of noncarbonate clay- and silt-size impurities and no quartz sand. If these impurities do exist, no problem would arise in expanding the reservoir chart to include impure and sandy facies. In addition, two special rock types are included. These, although easily placed in the reservoir chart on the basis of their original sedimentary texture, have suffered unique physico-chemical alteration, probably leaching, which has resulted in pinpoint porosity and voids in clastic carbonate fragments.

Another important addition to the reservoir-rock chart is that numbers and letters have been assigned to rock types corresponding to the various criteria used in their classification. This permits the processing of core data by IBM methods. By using a maximum of two digits and two letters it is possible to designate (1) original sedimentary texture, (2) degree of alteration in original sedimentary texture, (3) type of alteration, and (4) amount of visual porosity. An explanation of the meaning of each letter and numeral is given below.

Characteristics which distinguish the various carbonate rock types have been described in the general classification. As these elements of general classification are equally applicable to the special group of carbonates—the reservoir rocks—only exceptions will be repeated here.

Original Sedimentary Texture

Fine-grained limestone, calcarenitic limestone, calcarenite, and coarse carbonate clastic, subdivided on original sedimentary textural differences, have been designated groups 1, 2, 3, and 5 respectively. Thus the first numeral shows whether the rock consisted originally of particles of clay-silt, sandy mud, sand, or conglomerate size.

Degree of Alteration in Original Texture

The second numeral has been used to indicate the degree to which the original sedimentary texture has been altered. Four basic steps are recognized in which original texture is: (1) not visibly altered, (2) moderately altered, (3) strongly altered, and (4) obliterated.

Original texture not visibly altered.—Rocks of groups 1, 2, 3, and 5 that have remained essentially unaltered (except for compaction and cementation) are respectively termed Type 1,

TABLE 2.—ARABIAN CARBONATE RESERVOIR-ROCK CLASSIFICATION

	ORIG. TEXTURE NOT VISIBLY ALTERED (except by cementation)	ORIGINAL TEXTURE ALTERED					ORIGINAL TEXTURE OBLITERATED
		MODERATELY		STRONGLY			
	LESS THAN 10% DOLOMITE	LESS THAN 10% DOLOMITE	MORE THAN 10% PORPHYROBLASTIC DOLOMITE	LESS THAN 10% DOLOMITE	**10-75% DOLOMITE	MORE THAN 75% DOLOMITE WITH RELIC TEXTURE	MORE THAN 75% DOLOMITE
GROUP 1 *FINE-GRAINED LIMESTONE* Fine-grained limestone with less than 10% sand size or coarser than sand size clastic carbonate fragments*	1A, 1B, 1C FINE-GRAINED LIMESTONE	11AR, 11BR, 11CR PARTIALLY RECRYSTALLIZED, FINE-GRAINED LIMESTONE	11AD, 11BD, 11CD PARTIALLY DOLOMITIZED, FINE-GRAINED LIMESTONE	12AR, 12BR, 12CR CRYSTALLINE LIMESTONE	12AD, 12BD, 12CD DOLOMITIC LIMESTONE	13A, 13B, 13C CALCITIC DOLOMITE	
GROUP 2 *CALCARENITIC LIMESTONE* More than 10% sand size or coarser than sand size clastic carbonate fragments* in more than 10% original fine-grained matrix	2A, 2B, 2C CALCARENITIC LIMESTONE	21AR, 21BR, 21CR PARTIALLY RECRYSTALLIZED, CALCARENITIC LIMESTONE	21AD, 21BD, 21CD PARTIALLY DOLOMITIZED, CALCARENITIC LIMESTONE	22AR, 22BR, 22CR CRYSTALLINE, CALCARENITIC LIMESTONE	22AD, 22BD, 22CD DOLOMITIC, CALCARENITIC LIMESTONE	23A, 23B, 23C CALCARENITIC DOLOMITE	**GROUP 4** 4A, 4B, 4C CRYSTALLINE DOLOMITE
GROUP 3 *CALCARENITE* Sand size clastic carbonate fragments* in less than 10% original fine-grained matrix	3A, 3B, 3C CALCARENITE	31AR, 31BR, 31CR PARTIALLY RECRYSTALLIZED CALCARENITE	31AD, 31BD, 31CD PARTIALLY DOLOMITIZED CALCARENITE	32AR, 32BR, 32CR STRONGLY RECRYSTALLIZED CALCARENITE	32AD, 32BD; 32CD DOLOMITIC CALCARENITE	33A, 33B, 33C CALCARENITE DOLOMITE	
GROUP 5 *COARSE CLASTIC CARBONATE* Coarser than sand size clastic carbonate fragments* in less than 10% original fine-grained matrix	5A, 5B, 5C COARSE CLASTIC CARBONATE	51AR, 51BR, 51CR PARTIALLY RECRYSTALLIZED, COARSE CLASTIC CARBONATE	51AD, 51BD, 51CD PARTIALLY DOLOMITIZED, COARSE CLASTIC CARBONATE	52AR, 52BR, 52CR STRONGLY RECRYSTALLIZED, COARSE CLASTIC CARBONATE	52AD, 52BD, 52CD DOLOMITIC, COARSE CLASTIC CARBONATE	53A, 53B, 53C COARSE CLASTIC CARBONATE DOLOMITE	

		CARBONATE	CARBONATE	CARBONATE			

SPECIAL ROCK TYPES

(Not readily placed in reservoir rock classification as a result of unique physicochemical alteration.)

GROUP 1	14AX FINE-GRAINED LIMESTONE WITH SMALL VISIBLE VOIDS
	14BX COMPACT, FINE-GRAINED LIMESTONE WITH SMALL VISIBLE VOIDS

GROUP 3	31BX STRONGLY RECRYSTALLIZED CALCARENITE (VOIDS IN PLACE OF OOLITHS)
GROUP 5	51BX STRONGLY RECRYSTALLIZED COARSE CLASTIC CARBONATE (VOIDS IN PLACE OF FRAGMENTS)

* Clastic carbonate fragments include sand size and coarser than sand size grains irrespective of origin, for example, particles of older carbonate rock, ooliths, pellets and biogenic trash.

** Dolomitic and calcitic portions of rock so intimately intergrown as to be generally visually indistinguishable.

KEY TO RESERVOIR ROCK CLASSIFICATION SYMBOLS

1st Numeral - Designates rock groups. Groups 1,2,3, & 5 distinguished on basis of original textural differences; Group 4 on complete alteration of any of these to crystalline dolomite.

2nd Numeral - Indicates degree of alteration in original texture, excluding compaction and cementation. 1 = Moderately altered original texture. 2 = Strongly altered original texture. 3 = Relic original texture.

1st Letter--- Shows amount of visual porosity. A = Porous and/or poorly cemented. B = Compact and/or moderately cemented. C = Dense and/or strongly cemented.

2nd Letter--- Type of alteration. R = Recrystallization involving rearrangement of materials already present. D, A, S = Recrystallization of original sediment brought about by introduction and/or replacement by dolomite, anhydrite or silica respectively. X = Alteration brought about by incompletely understood physicochemical changes.

Mil --------- Millepore

2, 3, and 5. The lack of alteration is indicated by an implied (0), for exmple: 1(0), 2(0), etc.

Original texture moderately altered.—The carbonate rocks which have been subjected to moderate alteration (partial recrystallization and partial dolomitization) are designated Type 11, 21, 31, and 51 depending on their group affinities.

Original texture strongly altered.—Crystalline, strongly recrystallized, and dolomitic alteration are indicated by Types 12, 22, 32, and 52, whereas relic original texture caused by intense dolomitization is represented by 13, 23, 33, and 53 respectively.

Original texture obliterated.—When the original texture of any Group 1, 2, 3, and 5 rock has been completely altered to crystalline dolomite, the rock is typed as Group 4.

Type of Alteration

An explanation of the use of the first letter is given in the section on porosity; the second letter indicates the type of alteration involved.

R = Recrystallization by rearrangement of existing calcite without addition of appreciable dolomite

D = Recrystallization brought about by addition and/or replacement by dolomite

X = Alteration brought about by incompletely understood physico-chemical changes

Other letters such as A, S, etc., are used to indicate recrystallization involving anhydrite, silica, etc., respectively.

Amount of Visual Porosity

The final step in the classification of a reservoir rock is the division of each type on the basis of relative pore space. The porosity of rocks can be judged only approximately by visual inspection, and rarely can secondary porosity be seen in cuttings. Primary porosity, however, can be seen and estimated. The degree of accuracy obtained depends largely on experience.

The first letter in the rock type symbol is used to indicate the amount of visual porosity: A = porous and/or poorly cemented; B = compact and/or moderately cemented; and C = dense and/or strongly cemented. A separate treatment of each type and the resulting subtypes is not undertaken here. The primary objective will be to outline the main criteria used to distinguish these subtypes. These criteria are similarly applicable to all but a few reservoir rock types. The exceptions will be pointed out.

Normal or "A" porosity.—Rocks with normal porosity have generally undergone little alteration of original porosity. These rocks have large amounts of space between the grains or crystals, usually easily visible under the binocular microscope or hand lens. Rocks of highest porosity are generally well sorted, and the particles are rounded, or the crystals are widely spaced, joined mostly by the apices. The amount of pore space they contain is roughly equivalent to that of a moderately to highly porous sandstone.

Because of the wide variation in the size of carbonate rock particles and in textural patterns, the void space is at times not readily visible. This is particularly true of fine-grained limestone and calcarenitic limestone. The nature of very fine-grained limestone is such that the actual voids may not be seen even with the aid of a binocular microscope. In this case the over-all textural appearance of the rock sample should be observed. The fine-grained limestone with normal porosity generally has a distinct, easy to recognize, "chalky" texture. A drop of water applied to a dry surface will be quickly absorbed. Another simple way to estimate the porosity of fine-grained limestone is by comparable weight, provided large enough samples are available. If the weight does not noticeably vary from that of a piece of calcite of equal volume, then the specimen cannot be considered porous; if it is noticeably lighter then the decrease in weight is probably due to voids, and the rock is considered porous.

In calcarenite and coarse carbonate clastic, the voids are large and readily visible. The amount of void space is usually determined by the amount of cement between grains. For this reason, poorly cemented calcarenite and coarse carbonate clastic are the most porous. In most cases these rocks are friable to semi-friable. When dolomitization has resulted in replacement of all the calcite, then a highly porous rock sometimes results which is made up of loosely intergrown dolomite euhedra. A general rule is, if upon first glance through the binocular microscope the pore space is readily visible, then the rock has normal porosity.

Limited or "B" porosity.—Rocks with limited porosity have spaces between the grains or crystals which may be filled in part by matrix or cement with some visible pore space. This type of porosity is intermediate between

normal or "original" porosity and negligible porosity. It is by comparison with these two end members that an observer can best calibrate the visual criteria used to distinguish limited porosity. Rocks with limited porosity are described as compact or moderately cemented.

When lifted, they weigh approximately the same as a dense carbonate. Rocks with limited porosity are obviously more compact than the porous rocks and are usually more strongly cemented. A compact, very fine-grained limestone does not have the "chalky" texture so apparent in its porous equivalent. A drop of water placed on a dry surface will be absorbed at a much slower rate. Dolomites falling into this category are made up of a mosaic of finely to coarsely crystalline dolomite with occasional voids or vugs, usual formed by incomplete intermeshing of dolomite euhedra.

Negligible or "C" porosity.—When porosity has been reduced to insignificant amounts through any of several processes—*i.e.*, compaction and cementation—the rock is said to have negligible porosity and is described as dense. For a rock to be placed in this subtype, it must generally lack visual pore space. Dense, fine-grained limestone is often lithographic and has a characteristic conchoidal to subconchoidal fracture. In clastic carbonate rocks with negligible porosity all or nearly all the space between grains is filled with cement and/or matrix.

Special Rock Types

Several rock types have undergone unique physico-chemical alterations and as a consequence are not readily placed as to subtype in the reservoir rock classification. However, they are readily assigned to a group, and degree of alteration is apparent.

The first of these special types is fine-grained limestone which has undergone moderate alteration, possibly by leaching. The end result is a fine-grained limestone with small visible voids (pinpoint porosity) not related to primary texture. Rocks of this kind, which are relatively uncommon, are subtyped 14AX or or 14BX depending upon the amount of porosity.

The second, and more common special rock type, is calcarenite or coarse carbonate clastic which has voids in place of oöliths, presumably as a result of solution. The texture is strongly altered. The rocks are called strongly recrystallized calcarenite or coarse carbonate clastic and subtypes 31BX and 51BX respectively.

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LOCATION OF THE VALLEY HEADS MORAINE NEAR TULLY CENTER, NEW YORK, DETERMINED BY PREGLACIAL DIVIDE

BY FORREST DURHAM

The location of the Valley Heads Moraine, extending for many miles across New York, and the most conspicuous accumulation of drift in the State, may yield important clues to preglacial geomorphology.

The location of that portion of the moraine that blocks the valley near Tully Center, New York, and is referred to as the Tully Moraine, is structurally controlled. Its site is determined by the steep preglacial divide; the morainic material is plastered against the divide. Similar moraines in near-by valleys are controlled in like fashion.

Von Engeln (1919) considered the moraine an in-valley divide. By this he implied a deep trough extending south of Tully Center.

Fairchild (1925) mapped the preglacial divide near South Cortland. He also visualized a deep trough in the valley from Tully Center to Cortland filled with glacial outwash.

MacClintock and Apfel (1944) describe much of the Valley Heads as "valley stopper". They recognized that ice tongues lay north of the bedrock divides for a great length of time. However, they did not recognize the possibility that the moraine might be close to the preglacial divide. They suggested that some valley trains sloping southward from the moraines may cross the preglacial divides. The valley train south of the moraine near Tully Center extends about 12 miles to Cortland.

A number of factors argue against the preglacial divide being as far south as Cortland. As von Engeln noted, there is a definite dendritic pattern in both the Onondaga and Cortland valleys and their tributaries. This pattern is apparently a remnant of old consequent drainage from Tertiary time. Any extensive stream reversal should be reflected in the patterns of the tributary streams.

South of Tully Center the land surface slopes southward. There are four conspicuous cols between the east and west branches of the Tioughnioga River from approximately Tully Lake to Cortland. These cols in order from north to south have elevations of 1510, 1310, 1270, 1140 feet. Each is at the head of a tributary stream. Had the preglacial stream been flowing northward from Cortland toward Syracuse, the elevations of the tributaries should decrease northward.

Data from well drillers were assembled and plotted (Table 1). The most northerly well record in the valley was at Little York Lake where rock was encountered at a depth of 220 feet. A seismic crew made determinations at Tully Lake and at the junction of the side valley from Apulia with the main valley at Tully Center. Records show bedrock at Tully Lake at a depth of approximately 225 feet, and at 86 feet at the valley junction. It is suggested that the preglacial divide was at the location of the Tully Moraine (Fig. 2).

TABLE 1.—SUBSURFACE DATA

Location	Elevation	Depth to rock	Elevation of rock
Fabius	1285	70	1215
Tully Center	1240	30	1210
Junction of Valley from Fabius	1225	86*	1139
Tully Lake	1200	225*	975
Near Bromley (Otisco Valley)	1320	352	968
Little York	1160	220	940
South Cortland	1200	300	900
Truxton	1150	16	1134
1 mile northeast of Cortland (East Branch)	1100	50	1050
1 mile east-southeast of Cortland	1100	80	1020
Freeville	1020	165	855

* Denotes Seismic Data.

Bedrock at Bromley, approximately at the position of the Valley Heads in the Otisco Valley, is at a depth of 352 feet, or at an elevation of 968 feet above sea level. The surface elevation of Otisco Lake is 788 feet. The preglacial divide in the Otisco Valley is also at or near the present divide.

The findings in these valleys suggest corre-

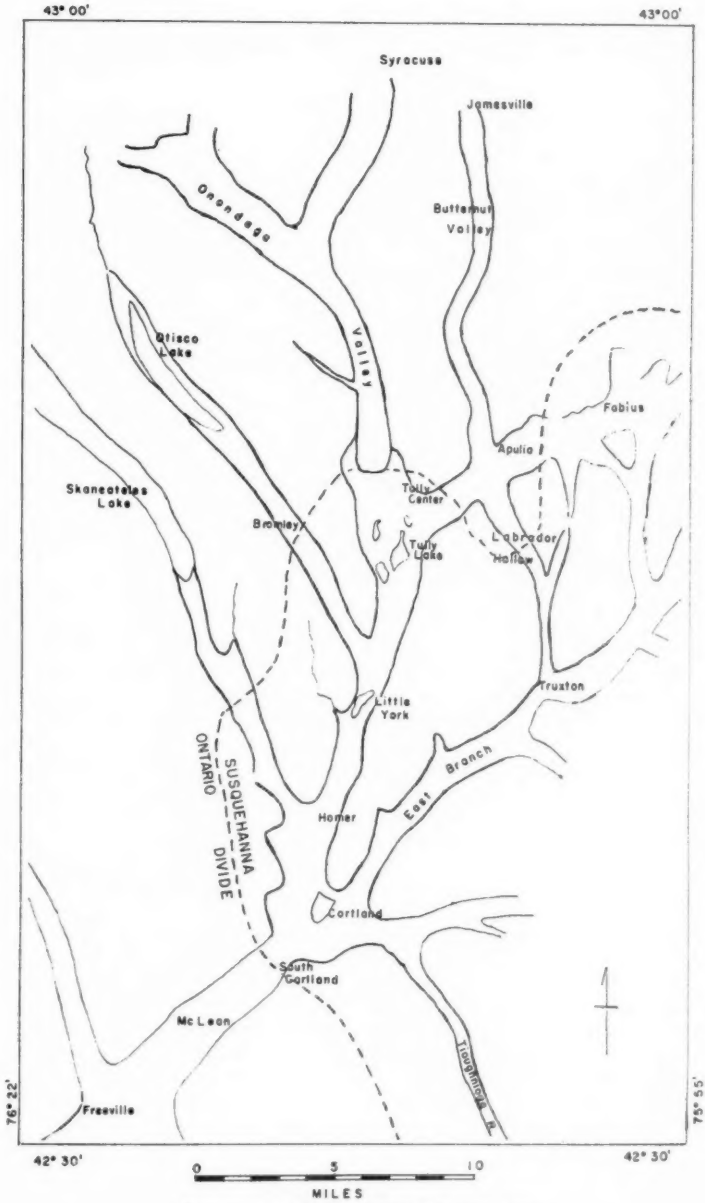

FIGURE 1.—MAP OF TULLY MORAINE AREA

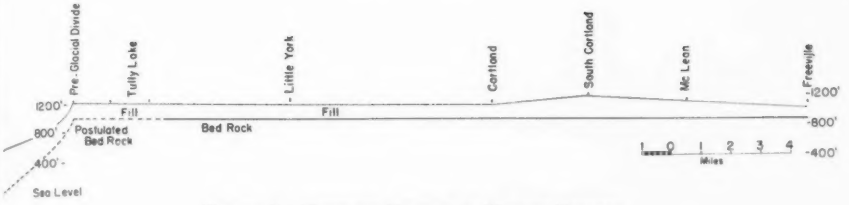

FIGURE 2.—BEDROCK PROFILE IN VALLEY BOTTOM

sponding conditions in the several other portions of the Valley Heads Moraine across the State.

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